Long-term effects of soil organic matter burial on carbon sequestration

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Herzlich bedanke ich mich bei. . .

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Abstract

Soils represent the largest terrestrial carbon reservoir. Preservation and additional accumulation of soil organic carbon (SOC) are of great interest to scientists and politicians as a measure of climate change mitigation and adaptation because of the key role of SOC in the global C cycle. Most SOC sequestration measures, such as no-tillage, agroforestry and application of animal manure focus on enhancing SOC stocks in the topsoil, which is regularly maximum 30 cm deep. However, 53% of SOC down to 100 cm is stored below 30 cm depth. It has been widely suggested that subsoils hold a larger potential to store additional OC than topsoils because of the large number of unsaturated mineral surfaces and environmental conditions that slower SOC decomposition. Burial of C-rich soil material, e.g. by colluvial deposition, generally leads to a long-term increase of landscape scale SOC stocks. In this dissertation, the option of actively burying SOC-rich soil material through tillage practices was studied for the first time assessing SOC sequestration in deep ploughed arable and forest soils as well as in fossil ridge and furrow cropland under forest with buried topsoils.

In the first part of this thesis, we sampled five loamy and five sandy cropland soils as well as four sandy forest soils that were ploughed to 55-90 cm depth 25-48 years ago. Adjacent, similarly managed but non-deep ploughed plots were sampled as reference. The deep-ploughed cropland soils contained on average 42±18% more SOC than the reference plots down to 100 cm. In contrast, total SOC stocks in deep ploughed forest soils were not significantly higher than in their respective reference soils (103±11 and 105±9 Mg ha^{-1}, respectively, p=0.2). However, SOC stocks in subsoils of deep ploughed plots were higher than in reference plots both under forest and under cropland. The 'newly formed' topsoil of the deep ploughed soils still contained less SOC than the reference topsoil, 14% and 37% under cropland and under forest, respectively. This was attributed to the slow SOC accumulation in deep ploughed forest topsoils, partly related to N limitation and acidic conditions.

Buried SOC stability was assessed through one-year incubation experiments, which enabled a comparison of SOC turnover in buried topsoil stripes and reference topsoils under standardized laboratory conditions eliminating possible temperature, oxygen or water limitations. On average, the fraction of mineralised SOC, was 32% lower in incubated buried than in reference topsoils. Sandy cropland buried topsoil stripes and reference topsoils had the lowest mineralisable SOC fraction (27±4 and 40±6 mg CO_2-C g^{-1} SOC, respectively). Buried SOC was partly to fully preserved, most effectively in sandy cropland soils, but selective preservation of certain, most stabile SOC fractions and preferential decomposition of labile fractions, such as the commonly most easily decomposable free light fraction, could not explain SOC stability. Land use history of many of the studied sandy sites as heath was identified as an important aspect concerning SOC stabilisation.

In the second part of this thesis, the impact of medieval ridge and furrow cropland management on SOC in five forest soils of different texture was assessed. Continuous clockwise ploughing in rectangular fields moved topsoil from the outer part of strip-shaped fields towards the centre,

thus forming a corrugated microtopography with peripheral furrows and central ridges. This tillage technique led to the burial of former topsoil under the ridges. SOC content was 0.4-0.9 g kg^{-1} higher at ridges than at reference positions, indicating long-term preservation of former topsoil SOC by burial under ridges, although enhanced SOC stocks at ridges due to carbon burial could not be significantly confirmed for all sites. Another important factor in these systems was the modified C input patterns in relation to the microtopography. Enhanced C input at furrows through leaf litter accumulation was indicated by higher SOC content in the free light fraction (10$\pm$5 and 6$\pm$3 g kg^{-1}, respectively) and higher specific SOC mineralisation at furrows than at ridges (37$\pm$4 and 31$\pm$3 µg CO_2-C g^{-1} SOC, respectively).

In conclusion, deep ploughing led to enhanced subsoil SOC contents and stocks compared to a reference plot in forests and cropland. Deep ploughing increased the storage space for SOC by mechanically translocating topsoil material into the subsoil, where originally relatively less SOC is stored in most soil types. However, the establishment of new topsoil horizons that continue to accumulate SOC even more than four decades after deep ploughing additionally determine the effect of deep ploughing on full profile SOC stocks. Furthermore, deep ploughing may enhance the rooting depth and water storage capacity and thus provide less fluctuant stabilise plant growth conditions under climate change conditions. In forests on land with historical use of a ridge and furrow system, recent processes with spatial heterogeneous C input due to preserved microtopography determine SOC distribution, but not necessarily also SOC stocks. Buried topsoil could be detected even centuries after burial.

Kurzfassung

Böden stellen durch die Speicherung von organischer Bodensubstanz (OBS) den größten terrestrischen Kohlenstoffspeicher dar. Aufgrund der Schlüsselrolle der OBS im globalen Kohlenstoffkreislauf, hat die Erhaltung und Erhöhung des OBS-Vorrats eine hohe Relevanz als Klimaschutzmaßnahme und im Rahmen der Anpassung an den Klimawandel. Die meisten Kohlenstoffsequestrierungsmaßnahmen im Landmanagement, z.B. die pfluglose Bodenbearbeitung, die organische Düngung oder die Einführung von mehrjähringen Pflanzen durch Agroforstwirtschaft, konzentrieren sich auf eine Erhöhung der OBS-Vorräte in den obersten 30 cm des Bodens. Allerdings sind 53% der OBS bis 100 cm Tiefe unterhalb von 30 cm gespeichert. Aufgrund der hohen Anzahl an ungesättigten Mineraloberflächen sowie der Umweltbedingungen, welche die Mineralisierung der OBS verlangsam, betonen zahlreiche Studien, dass Unterböden ein größeres Kohlenstoffsequestrierungspotential aufweisen als Oberböden. Die Vergrabung von OBS-reichen Bodenmaterial durch beispielsweise kolluviale Ablagerungen führt generell zur Landzeiterhöhung von Kohlenstoffvorräten auf der Landschaftsskala. In dieser Dissertation wurde zum ersten Mal die Option der aktiven Vergrabung von OBS-reichem Material durch Landmanagementpraktiken hinsichtlich Ihrer Kohlenstoffsequestrierungsleistung untersucht. Insbesondere wurden tiefgepflügte Acker- und Waldböden sowie fossile Wölbackerfluren unter Wald untersucht.

Im ersten Teil der Arbeit wurden jeweils fünf lehmige und sandige Ackerböden sowie vier sandige Waldböden untersucht, die vor 25-48 Jahren 55-90 cm tief gepflügt wurden. Benachbarte, ähnlich bewirtschaftete jedoch nicht tiefgepflügte Teilflächen wurden als Referenz beprobt. Die tiefgepflügten Ackerböden enthielten durchschnittlich 42±18% höhere Kohlenstoffvorräte als die jeweiligen Referenzböden bis in 100 cm Tiefe. Im Gegensatz waren die Kohlenstoffvorräte bis in 100 cm Tiefe in tiefgepflügten Wälder nicht signifikant höher als die der jeweiligen Referenz (103±11 and 105±9 Mg ha^{-1}, jeweils, $p=0.2$). Im Unterboden jedoch waren die Kohlenstoffvorräte der tiefgepflügten Böden höher als die der Referenz sowohl unter Acker als auch unter Wald. Der 'neu etablierte' Oberboden in den tiefgepflügten Böden hatte 14% bzw. 37% niedrigere Kohlenstoffvorräte als die Referenz jeweils unter Acker und unter Wald. Dieses Defizit wurde dadurch erklärt, dass OBS in den tiefgepflügten Waldoberböden sich langsamer akkumuliert im Vergleich zu Ackeroberböden bzw. Ackerkrumen, zum Teil aufgrund von sauren pH-Werten und stickstoffarmen Bedingungen.

Die Stabilität der vergrabenen OBS wurde anhand einjähriger Inkubationsexperimente untersucht. Diese ermöglichten, die Umsetzung der OBS im vergrabenen mit der im Referenzoberboden unter standardisierte Laborbedingungen zu vergleichen. Dabei wurden mögliche Temperatur-, Sauerstoff- und Wasserlimitierungen für die Umsetzung ausgeschaltet. Durchschnittlich was der Anteil an während der Inkubation mineralisierter OBS 32% niedriger in den vergrabenen Oberböden im Vergleich zur Referenz. Die sandigen Ackerböden wiesen die niedrigste mineralisierbare OBS-Fraktion auf, sowohl die vergrabenen als auch die Referenzoberböden (27±4 and 40±6 mg CO_2-C g^{-1} SOC, jeweils). Die OBS wurde seit ihrer Vergrabung teilweise komplett erhalten, am effektivsten in sandigen Ackerböden. Selektive Erhaltung bestimmter OBS-Fraktionen mit

höherer Stabilität und gleichzeitiger Abbau labiler Fraktionen wie die freie leichte Fraktion, konnte die hoher OBS-Stabilität nicht erklären. Als ein wichtiger Aspekt im Zusammenhang mit der hohen OBS-Stabilität, wurde die Geschichte einer Landnutzung als Heide bzw. die Moorvergangenheit vieler der untersuchten Standorte identifiziert.

Im zweiten Teil dieser Arbeit wurde die Auswirkung von mittelalterlicher Wölbackerbewirtschaftung auf die OBS in fünf Waldböden mit unterschiedlicher Textur untersucht. Regelmäßiges, spiralförmiges Pflügen in länglichen Parzellen führte dazu, dass sich Oberboden aus dem Rand streifenförmiger Felder in deren Mitte akkumulierte. Auf diese Art und Weise entstand eine gewölbte Mikrotopographie mit Furchen in den Außenbereichen und Kämmen in der Mitte. Durch diese altertümliche Bearbeitungstechnik wurde ehemaliger Oberboden unter den Kämmen vergraben. Der Kohlenstoffgehalt war 0.4-0.9 g kg^{-1} höher unter Kämmen als an einer Referenzposition. Dies deutet auf eine Langzeiterhaltung der OBS im ehemaligen Oberboden unter den Kämmen hin. Allerdings konnten erhöhte Kohlenstoffvorräte unter den Kämmen aufgrund von Oberbodenvergrabung nicht für alle Standorte signifikant bestätigt werden. Eine wichtige Rolle in diesen Systemen spielte auch die Veränderung des Kohlenstoffeintrags durch die entstandene Mikrotopographie. Indikatoren für einen höheren Kohlenstoffeintrag in den Furchen durch Laubanreicherung waren die höheren Anteile an freier leichter OBS-Fraktion (10 ± 5 and 6 ± 3 g kg^{-1}, jeweils) sowie die höhere spezifische Mineralisation der OBS in Oberböden der Furchen als in Oberböden der Kämmen (37 ± 4 and 31 ± 3 µg CO_2-C g^{-1} SOC, jeweils).

Zusammenfassend lässt sich schlussfolgern, dass Tiefpflügen zu einer Erhöhung von Kohlenstoffgehalten und -vorräten in Unterböden unter Wald und Acker geführt hat. In tiefgepflügten Böden wurde der Speicherraum für Kohlenstoff durch die mechanische Translokation von Oberbodenmaterial in größeren Tiefen erhöht. Allerdings wurden dabei auch neue Oberböden etabliert, in den sich selbst über 40 Jahre nach dem Tiefpflügen Kohlenstoff weiter akkumuliert. Dieser Prozess bestimmt zusätzlich den Kohlenstoffsequestrierungseffekt bei der Betrachtung von Gesamtprofilvorräten. Darüber hinaus kann Tiefpflügen höhere Durchwurzelungstiefen begünstigen sowie Wasserhaltekapazitäten im Unterboden erhöhen. Dies könnte weniger schwankende Pflanzen- wachstumsbedingungen bedeuten, die unter Klimawandelszenarien eine wichtige Rolle spielen werden. In Wälder mit historischer Wölbackernutzung war der räumlich heterogene Kohlenstoffeintrag bedingt durch die erhaltene Mikrotopographie entscheidend für die Kohlenstoffverteilung im System. Dennoch konnte vergrabener Oberboden durch hohe Kohlenstoffgehalte Jahrhunderte nach der Vergrabung nachgewiesen werden.

1 Introduction

1.1 Relation of climate change to soil organic carbon

The global carbon (C) cycle is one of the main element cycles that influence the Earth's climate. About half of the incoming solar energy into the atmosphere is absorbed by the Earth's surface (Cubasch et al., 2013). Radiation reflected and emitted by the surface is largely absorbed and again reflected by gases such as water vapour, carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O). Atmospheric CO_2 represents hereby the main atmospheric phase of the global C cycle (Ciais et al., 2013). The part of the radiation that is redirected toward the Earth's surface through atmospheric gases additionally heats it and the lowest layers of the atmosphere up (greenhouse effect). An increase in the concentration, i. e. a perturbation of the natural C cycle, of the mentioned greenhouse gases can thus substantially enhance the reflecting radiation and thus Earth's surface temperatures.

Since the beginning of the industrial Era in 1750, concentration of greenhouse gases in the atmosphere has increased continuously, mostly due to human activites (D. J. Hartmann et al., 2013). For example, atmospheric CO_2 concentrations fluctuated between 180 and 290 ppm for 2.1 million years prior to the industrial Era (Hönisch et al., 2009; Petit et al., 1999). In 2014, atmospheric CO_2 abundance reached over 397 ppm (Le Quéré et al., 2015) and was thereby 40% higher than before industrialisation. On a cumulative basis, atmospheric CO_2 increase between 1750 and 2011 was 240±10 Pg C. Main anthropogenic sources for greenhouse gas emissions are fossil fuel combustion and oxidation, cement production, land use and land use change (Le Quéré et al., 2015). It is generally well accepted that increasing greenhouse gases concentrations have a direct connection to increasing surface temperatures, which have been observed over the last century. The change in atmospheric composition and the related temperature increases have manifold consequences for Earth's climate and is usually summarised as 'climate change' (Planton, 2013).

The emission of greenhouse gases like CO_2 from sediments (e.g. via fossil fuel combustion) or a terrestrial reservoir (e.g. via land use) represents, however, only one part of the global C cycle. Vegetation uptakes CO_2 from the atmosphere through photosynthesis. Dead biomass is not completely decomposed releasing all fixed C again to the atmosphere, but is partly transferred into the soil as litter, which over time accumulates to soil organic matter (SOM). The C contained in SOM is commonly referred to as soil organic carbon (SOC). This explains why soils represent the largest terrestrial reservoir of C within the global C cycle. While until 2011, 829 Pg C were in the atmosphere, living biomass contained 450 to 650 Pg C and soils 1500 to 2400 Pg C (Ciais et al., 2013).

The amount of C stored in soils is steered by the balance of C entering the soil, mainly via plants, and the C return to the atmosphere or transfer into other reservoirs such as rivers and oceans. The amount of C release to the atmosphere is mainly dependant on heterotrophic respiration, i.e.

decomposition of dead biomass or SOM to its mineral components such as CO_2 (Planton, 2013). Thus, there is a direct connection between the amount of SOM stored in soils and atmospheric CO_2 concentration.

1.2 Influence of agriculture and land management on soil organic carbon

Human activity has been modifying ecosystems since the Neolithic Age through land management and land use changes (Price, 2000). These have affected the global carbon (C) cycle through either C losses to the atmosphere or C sequestration in biomass and soils. By late medieval times, 4.6 x 106 km^2 land world-wide, equivalent to 5% of the area potentially covered by vegetation, was under agricultural use (Pongratz et al., 2008) and atmospheric CO_2 exceeded its natural range of variation (Pongratz et al., 2009). Between 1750 and 2011, one-third of all anthropogenic greenhouse gases was derived from land use changes (D. J. Hartmann et al., 2013).

Cultivation of native land leads to a SOC loss of approximately 30% (Davidson and Ackerman, 1993) because of the reduced plant derived C input, which results from clearance of the former vegetation cover and agricultural harvesting. As a result of land use change and soil cultivation, more C was emitted from soils to the atmosphere than the amount originated from fossil fuel combustion, until the 1950s (Lal, 2004b). Due to population growth, agricultural land continues increasing in most regions of the world (Kauppi et al., 2001). Between 1961 and 2002, almost 500 million hectares were converted to agriculture from other land uses worldwide (Nieder and Benbi, 2008). Conversion of grassland or forest to cropland leads to a SOC loss of around 34% reaching an equilibrium approximately 20 years after land use change (Poeplau et al., 2011). Inversely, afforestation of former cropland can lead to a 25-53% SOC accumulation, depending on the climate zone (Guo and Gifford, 2002; Laganière et al., 2010).

Apart from land use conversions such as afforestation, SOC accrual can be achieved through the implementation of certain land management practices (Kauppi et al., 2001). The '4 per 1000 initiative' launched 2015 by the French government highlights the political interest to tap the potential of SOC depleted agricultural soils in mitigating climate change through enlargement and preservation of SOC stocks (Paustian et al., 2016). For agricultural, mineral soils of Europe, (Freibauer et al., 2004) identified the promotion of organic inputs on arable land, the introduction of perennials, i.e. grasses and trees on arable set-aside land (e.g. agroforestry), the promotion of organic farming, as most promising measures to sequester more SOC. Further recommended management practices to enhance SOC are to return crop residues to soil surface (mulch farming) (Lal, 2004b) and to include cover crops in cropping systems (Poeplau and Don, 2015).

One of the most frequently studied and most controversially discussed land management options in relation to climate change mitigation is the reduction or ceasing of tillage (no-till) (Powlson et al., 2014). A recent meta-analysis showed that within 15 years after adoption of no-till practices, on average 3.4 Mg SOC ha^{-1} were accumulated in the first 30 cm of the soil (Virto et al., 2012). However, this SOC accumulation effect has been shown to be limited to easily decomposable labile fractions (Six et al., 1999; Wander and Bidart, 2000; M. Kaiser et al., 2014) and to the upper part of the topsoil. Concomitantly, it is mostly outweighed by SOC losses at greater depth (D. Angers et al., 1997; Deen and Kataki, 2003; M. Carter, 2005; M. Kaiser et al., 2014). Moreover, tillage reduction reduces energy consumption and C emissions through the use of fossil fuels (Holland, 2004). This benefit may be, nevertheless, outweighed by the increased pesticide demand in no-till systems (Haddaway et al., 2016). Additionally, scientific evidence exists that N_2O emissions may rise under no-till because of the moister and denser soil conditions (Rochette et al., 2008).

The mentioned land management practices recommended to enhance C sequestration in soils focus mainly on the first cm of the soil, i.e. the topsoil. However, it has been observed that conventionally managed cropland soils with SOC stocks below the regularly ploughed layer higher than usual contain higher full profile SOC stocks than grassland with similar soil properties (Don et al., 2009). These findings point out that agricultural soils hold a greater potential to store SOC, namely in greater soil depth - in subsoils.

1.3 Organic carbon in subsoils

In most soil types, SOC contents, i.e. the mass of C per kg soil, decrease with increasing soil depth (Blume et al., 2010). However, because topsoils extend mostly not deeper than 30 cm and subsoils may reach several decimetres, mass of subsoils is generally many times higher than topsoil mass. Thus, even with a low SOC content, subsoils may store more OC than topsoils. Regarding whole soil profiles, global SOC stocks were estimated to be 684–724 Pg of C in the upper 30 cm, 1462–1548 Pg of C in the upper 100 cm, and 2376–2456 Pg of C in the upper 200 cm (Batjes, 2014). Thus, over 70% of the world's SOC is stored below 30 cm. Especially because of the low SOC contents of subsoils, their potential to store additional C has been emphasized (C. Rumpel et al., 2012). Beare et al. (2014) estimated that soils at greater depth have a higher capacity of storing additional C compared to topsoils because of a larger difference between the existing SOC content and the SOC saturation value, which the authors based on the mineral surface area.

Subsoil OC is characterised by high apparent radiocarbon (^{14}C) ages indicating mean residence times of C in the soil of several thousand years (Mathieu et al., 2015). This observation leads to the assumption that OC in subsoils is stabilized over the long-term. The currently largely debated factors that account for long-term stability of subsoil OC are:

- interaction with the mineral phase, such as clay and poorly crystalline minerals (Kleber et al., 2005), which protects SOC against oxidative destruction (Rumpel et al., 2011),
- low probability of a decomposer to meet an SOC particle due to low SOC concentration (Don et al., 2013), i.e. low microbial density together with low microbial diversity (Agnelli et al., 2004; Ekschmitt et al., 2008),
- physical protection of occluded particulate OC in aggregates and microaggregates (Moni et al., 2010; Rasmussen et al., 2005),
- low nutrient availability (Garcia-Pausas et al., 2008; Fierer et al., 2003),
- higher chemical recalcitrance of root derived C compared to litter derived C. Root input has been identified as one of the main sources for subsoil OC (Rasse et al., 2005),
- missing availability of fresh C with a high bioenergetic yield (Fontaine et al., 2007) and
- accumulation of chemically recalcitrant black C in subsoil horizons (Dai et al., 2005; Rumpel et al., 2009), among others.

Additionally, multiple C recycling through soil's microorganisms might result in high ^{14}C ages although the chemical or biological lability of SOM molecules is not necessarily low (Gleixner, 2013; Rethemeyer et al., 2005). Also, geogenic C from sediments such as loess might substantially decrease ^{14}C contents in soils (Angst et al., 2016b; Rumpel et al., 2011).

Lorenz and Lal (2005) emphasise that subsoils have the potential to store 760-1520 Pg additional C. At the same time, they point out that care should be taken when adding new C sources to subsoils because of the risk of enhanced mineralisation of present SOC. Nevertheless, increasing SOC stocks in subsoil is still recognised as promising means of substantial C sequestration in soils (C. Rumpel et al., 2012).

Key to enhancing SOC stocks in subsoils is an enhanced C input into greater soil depth. Naturally, main C sources to subsoils are plant roots and root exudates, dissolved organic matter and bioturbation (Rumpel et al., 2011). Recently, downward transport of particulate organic matter has been observed and pointed out as a C source to subsoils (Angst et al., 2016b). In volcanic soils, colloidal Fe/Al-humus complexes is an important process for C translocation into greater soil depth (Osher et al., 2003). In soils where clay migration down the soil profile (clay eluviation) is an important pedogenetic process, such as Alisols, Luvisols, Acrisols and Lixisols (IUSS Working Group WRB, 2015), C input into subsoils may occur as organomineral complexes (Rumpel et al., 2011). The promotion of deep rooting plants in managed systems as well as the activity of soil fauna that transports SOC downward in the soil profile have been suggested as measures to enhance C input into the subsoil (Lorenz and Lal, 2005).

An often unmentioned process that can redistribute large masses of C enriched topsoil material to a greater soil depth is SOC burial, i.e. at depositional landscape positions (Doetterl et al., 2015b). Between 70 and 90 % of the eroded topsoil material is deposited in the terrestrial landscape and not transported into lakes and oceans (Vente et al., 2007; Walling, 1996). This aspect of C transport to a greater soil depth, thus, requires attention within the discussion on SOC sequestration in subsoils.

1.4 Burial of carbon-rich soil material

In recent years, the role of soil redistribution within the landscape leading to burial of large amounts of SOC has received considerable attention (Johnson, 2014; K Van Oost et al., 2007). Environmental conditions in buried soils can largely differ from those near the surface because of a detachment from atmospheric conditions (N. T. Chaopricha et al., 2014).

Burial of SOC-rich material can occur via geomorphic, pedogenic and anthropogenic depositional processes. Beside the above mentioned soil erosion through wind and water with subsequent deposition at footslopes, colluvial valley bottoms and alluvial flood plains (Stallard, 1998), further important processes are:

- aeolian burial, i.e. gradual accumulation of loess deposits (Antoine et al., 2013; Mason et al., 2008),
- volcanic burial, i.e. covering under lava flows and other volcanic deposits, such as tephra, ashes or pumice (Basile-Doelsch et al., 2005; Zech, 2006),
- alluvial burial, i.e. covering by floodplain and alluvial sediments (Blazejewski et al., 2009; B. J. Carter et al., 2009),

- covering under glacial deposits (Harris et al., 1987; Mahaney et al., 2001),
- successive freezing and thawing events (cryoturbation) in permafrost environments (Becher et al., 2013; Bockheim, 2007) and
- covering after landslides and similar mass moving (mass wasting) processes (Mayer et al., 2008), among others.

Furthermore, buried SOC plays a role in relation to archaeological findings, such as burial mounds from the Bronze Age (Thomsen et al., 2008) or the similar 'Kurgans' in the steppe (Demkina et al., 2010), pre-Columbian raised fields in the Amazon Basin (Rodrigues et al., 2015) as well as medieval fortresses and ridge and furrow fields (Meibeyer, 1969).

Recent anthropogenic activity can additionally lead to an increase of SOC burial. Due to the missing plant cover and the disturbance of soil aggregates, conventionally ploughed fields have been estimated to have up to double as high erosion rates compared to soils under native vegetation (Montgomery, 2007). Alteration of river morphology can result in magnified flooding events and alluvial burial (Knox, 2006). Also, urban environments may contain large amounts of SOC under constructions (N. T. Chaopricha et al., 2014).

Through land management practices, C-rich soil material can also get transported to greater soil depth, although it is not necessarily their main purpose. Deep tillage includes several methods that entail the mechanical working of soil to a greater depth, comprising subsoiling, deep ploughing and deep mixing (Schneider et al., 2017). Subsoiling, also called deep ripping or deep chiseling, is an operation, which aims to loosen the soil structure and decrease the bulk density without critically modifying the soil profile. Deep ploughing turns soil horizons to a certain angle (often 130°) resulting in a soil profile consisting of alternating subsoil and topsoil stripes down to a greater soil depth (mostly deeper than 60 cm) than the common plough layer (Kuntze et al., 1994). Deep mixing techniques completely stir soil horizons, so that they cannot be distinguished afterwards.

Buried SOC has been observed to decompose more slowly than SOC in near-surface soil horizons. A 10,000 year old Brady Soil loess paleosol of the central Great Plains of the U.S. contained thermodynamically and biologically more stable SOC than that of the modern surface soil (Chaopricha, 2013). In Canadian depositional landscapes, buried SOC was less mineralisable in laboratory incubation than that found in surface soils (VandenBygaart et al., 2015). Wang et al. (2014) have calculated that around 17% of buried SOC by erosion is expected to be preserved 1000–1500 yr after burial.

1.5 Scope of the thesis

The yet untapped potential of subsoils to contribute to SOC sequestration in managed soils and thus play an important role in climate change mitigation has been presented. Burial of C-rich soil material has been often recognised as an important element of terrestrial C sequestration, although it is often disregarded in Earth System Models (Doetterl et al., 2015b).

This thesis was conducted with the aim of evaluating the option of active burial of large amounts of SOC, i.e. topsoil burial, by land management practices. Specifically, two practices were evaluated: deep ploughing and ridge and furrow cultivation. In both, topsoils were buried between

20 and 90 cm below the soil surface among the included study sites. Details on these land management practices can be found in Chapters 2 and 4, respectively.

The research for this thesis was conducted in the framework of the project "Burial of organic matter for carbon sequestration: Potentials, processes and long-term effects" funded by the Deutsche Forschungsgemeinschaft. The general aims of this project were:

1. to compare the long-term effect of topsoil burial on SOC stocks and thus, its climate impact, studying deep ploughed experimental sites under cropland and forest as well as forest sites with a historic land use as cropland, which involved topsoil burial.
2. to inquire the underlying mechanisms and processes of long-term stabilisation and storage of buried SOC analysing main soil specific driving factors that control C storage and the degree of SOC stabilisation

Chapter 2 represents the first quantification of the effect of deep ploughing on cropland SOC stocks. Insights into the stability of buried SOC are also provided here. Chapter 3 compares the impact of deep ploughing in arable and forest soils. The restoration of SOC stocks in newly formed topsoils of deep ploughed soils emerges as an important aspect to consider.

Because subsoil OC can be several thousand years old (Chabbi et al., 2009), SOC storage in subsoils also needed to be assessed at a larger timescale to be able to estimate how persistent is the effect of SOC burial. Chapter 4 presents the study of buried SOC under forests, which were used as arable ridge and furrow land during the middle ages.

1.6 Publications

Chapters 2-4 of this thesis correspond to the peer-reviewed publications listed below. Chapter 2 was published in the journal *Global Change Biology* in 2016. Chapter 3 has been accepted for publication in the journal *Scientific Reports*. Chapter 4 was published in the journal *Catena* in 2017.

Chapter 2:
Alcántara, V., Don, A., Well, R. and Nieder, R., (2016). Deep ploughing increases agricultural soil organic matter stocks. Glob. Chang. Biol. 22, 2939–2956. doi:10.1111/gcb.13289

Chapter 3:
Alcántara, V., Don, A., Vesterdal, V., Well, R. and Nieder, R. (accepted). Stability of buried carbon in deep ploughed forest and cropland soils - implications for carbon stocks. Scientific reports.

Chapter 4:
Alcántara, V., Don, A., Well, R. and Nieder, R. (2017). Legacy of medieval ridge and furrow cultivation on soil organic carbon distribution and stocks in forests. Catena. 154: 85-94, doi:10.1016/j.catena.2017.02.013.

2 Deep ploughing increases agricultural soil organic matter stocks

2.1 Abstract

Subsoils play an important role within the global C cycle, since they have high soil organic carbon (SOC) storage capacity due to generally low SOC contents. However, measures for enhancing SOC storage commonly focus on topsoils. This study assessed the long-term storage and stability of SOC in topsoils buried in arable subsoils by deep ploughing, a globally applied method for breaking up hard pans and improving soil structure to optimise crop growing conditions. One effect of deep ploughing is translocation of SOC formed near the surface into the subsoil, with concomitant mixing of SOC-poor subsoil material into the 'new' topsoil. deep-ploughed croplands represent unique long-term in situ incubations of SOC-rich material in subsoils. In this study, we sampled five loamy and five sandy soils that were ploughed to 55-90 cm depth 35-50 years ago. Adjacent, similarly managed but conventionally ploughed subplots were sampled as reference. The deep-ploughed soils contained on average 42±18% more SOC than the reference subplots. On average, 45 years after deep ploughing, the 'new' topsoil still contained 14% less SOC than the reference topsoil, indicating long-term SOC accumulation potential in the topsoil. In vitro incubation experiments on the buried sandy soils revealed 63±6% lower potential SOC mineralisation rate and also 67±2% lower SOC mineralisation per unit SOC in the buried topsoils than in the reference topsoils. Wider C/N ratio in the buried sandy topsoils than in the reference topsoils indicates that deep ploughing preserved SOC. The SOC mineralisation per unit SOC in the buried loamy topsoils was not significantly different from that in the reference topsoils. However, 56±4% of the initial SOC was preserved in the buried topsoils. It can be concluded that deep ploughing contributes to SOC sequestration by enlarging the storage space for SOC-rich material.

2.2 Introduction

Soils constitute the largest terrestrial carbon reservoir (Ciais et al., 2013) and can act as a sink or source for atmospheric CO_2. Thus, strategies to accomplish accumulation and long-term storage of soil organic matter (SOM) have become increasingly important in the context of climate change mitigation. Conservation and accumulation of SOM is also linked to soil fertility and food security, since SOM controls and improves major soil functions (Lal, 2004a). In arable soils these aspects are commonly attributed to properties of the regularly ploughed layer.

Conservation tillage (reduced tillage or no-till) is one of the most frequently studied options to enhance soil organic carbon (SOC) in croplands. However, in temperate climates such practices have not proven to be an effective climate change mitigation option (Powlson et al., 2014). Their SOC accumulation effect has been shown to be limited to easily decomposable labile fractions (Six et al., 1999; Wander and Bidart, 2000; M. Kaiser et al., 2014) and to the upper part of the topsoil, and is mostly outweighed by SOC losses at greater depth (D. Angers et al., 1997; Deen and Kataki, 2003; M. Carter, 2005; M. Kaiser et al., 2014).

Within the field of SOC sequestration research, attention has recently been drawn to subsoils, since they store approximately 53% of the SOC in the 0-100 cm layer and 71% between 0 and 200 cm depth (Batjes, 2014). In addition, subsoils have attracted interest because SOC radiocarbon age increases with depth (Gleixner, 2013) in all soil types (Mathieu et al., 2015), leading to the assumption that subsoil SOC is highly stable. At the same time, surprising evidence exists for the use of old SOC as an energy source for microorganisms (Fontaine et al., 2007; Gleixner, 2013). Consequently, in efforts to increase SOC sequestration by enhancing the C inputs into the subsoil, care should be taken to ensure that old stabilised SOC is not mineralised through the addition of easily decomposable compounds with so-called priming effects (C. Rumpel et al., 2012).

The importance of SOC burial as a SOC stabilisation mechanism has not been widely studied (N. T. Chaopricha et al., 2014). Processes such as deposition of eroded soil material and volcanic, aeolian or alluvial burial can transport larger SOC masses to greater depth in soil than the amounts derived from root litter or transported by dissolved OC and particulate organic matter. Burial of surface soils can lead to net accumulation of SOC (Kristof Van Oost et al., 2012; Hoffmann et al., 2013). However, buried SOC in depositional positions has been identified as only partly stabilised, and on a short-term basis (Doetterl et al., 2012; Doetterl et al., 2016). In contrast, the effects of SOC burial as an agricultural management option, i.e. through deep ploughing, have not been studied to date in the context of SOC sequestration.

Deep ploughing is carried out mostly only once, with the objective of improving soil functions such as water infiltration capacity and root penetration and thus optimising crop growing conditions. Other subsoil management techniques are performed more than once or even regularly, such as mixing of the soil profile with disc-type power cultivators, deep ripping, deep rototilling and deep discing. However, in this study we focus on deep ploughing conducted only once. Deep ploughing was first promoted in Europe after the invention of the steam plough in the late 19th century and enabled a ploughing depth of 40-50 cm (Eggelsmann, 1979; Russell, 2009). Only since the 1930s has deep ploughing been performed down to 60 to 120 cm depth and it is currently applied to many different agricultural soils worldwide. It has been widely used to cultivate heath soils by breaking up the hard pan of Podzols, thus enabling agricultural cultivation. Peatlands have also been cultivated by deep ploughing, on an area of almost 200 000 ha in Germany and in the Netherlands (Eggelsmann, 1979). In the USA, the effects of deep ploughing on crop growth have been studied since the 1960s (Burnett and Hauser, 1967). Although deep ploughing has become less common since the 1970s, it is still applied for hard pan or plough pan breakup, to enlarge the rooting zone or as a preparation measure for afforestation in countries such as Canada (Unger, 1979), Denmark (Hansen et al., 2007), the Netherlands (Ouwerkerk and Raats, 1986), Sweden (Nordborg and Nilsson, 2003), China (Cai et al., 2014) and the USA (Zobeck and Schillinger, 2010).

Deep ploughing inverts the soil using a mouldboard plough. The soil profile of cropland that has been deep-ploughed comprises diagonal stripes of alternating topsoil and subsoil material (Fig.

2.1). With continuous cultivation, a new topsoil horizon is established. Below the deep-ploughed stripes, the subsoil remains undisturbed. The new topsoil mixed with subsoil is expected to be able to store large amounts of SOC due to its lower SOC content (Chung et al., 2010) and high content of unsaturated mineral surfaces (Baldock and Skjemstad, 2000; Rasse et al., 2005). This has been reported e.g. for new topsoil of flipped soils (Thomas et al., 2007) and fresh slip scars in hilly areas (Rosser and Ross, 2011).

Regarding the question of increasing SOC sequestration in subsoil through tillage practices, here we studied the effects of burial of SOC-rich topsoil in the subsoil through deep ploughing. Field experiments with a paired subplot design established in the 1960s and 1970s, in which part of an agricultural field was deep-ploughed, were used in the study. The following hypotheses were tested:

1. Burial of topsoil material in the subsoil of arable soils through deep ploughing increases SOC and N stocks in the long term, irrespective of soil type and texture.
2. Long-term accumulation of SOC in the 'new' topsoil of deep-ploughed soils establishes SOC levels and stocks comparable to those in reference topsoil.
3. SOC in buried topsoil material is more stable than SOC in reference topsoil and SOC stability in buried topsoil material increases with soil depth.
4. The variability of the increase in SOC stocks through deep ploughing is related to soil properties such as pH, texture, cation exchange capacity and poorly crystalline Fe and Al

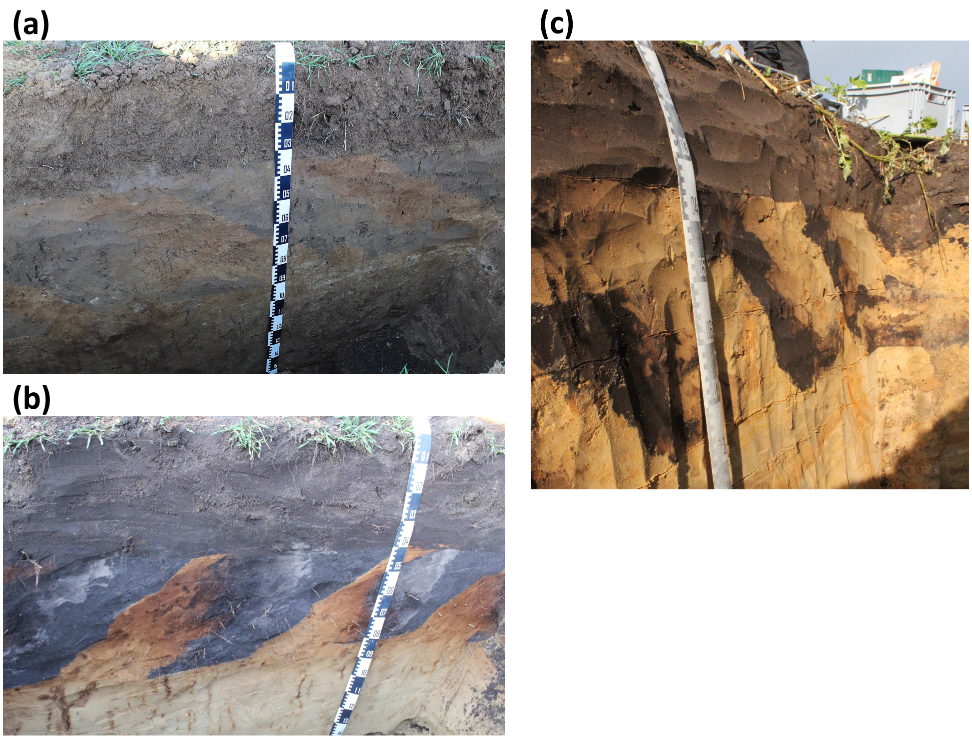

FIGURE 2.1: Deep-ploughed soil profiles **(a)** at loamy site SZ, **(b)** at sandy site HB and **(c)** at sandy site EM.

2.3 Materials and Methods

Sampling sites

We selected 10 field sites in which one part was deep-ploughed with a special mouldboard plough 36 to 48 years prior to sampling and the rest of the field was not deep-ploughed, serving as a reference. The sites were located in Northern and Central Germany. The field trials were set up by the local Chamber of Agriculture between 1965 and 1978 as practice trials on different farmers' field sites. Chemical and physical properties of the soils were determined by the Chamber of Agriculture prior to deep ploughing and in some cases also directly after deep ploughing, but no archived samples were available. Differences in yield between deep-ploughed plots and non-deep-ploughed reference plots were monitored for the initial 3-5 years.

Along the border of the deep-ploughed part of each site, we identified a representative plot for sampling and divided it into a deep-ploughed part and a directly adjacent reference subplot in the non-deep-ploughed part of the field site. Subplot size was 20 m by 40 m. At five of the sites the soil texture was loamy and at the other five sites it was sandy. The deep-ploughed subplots were ploughed to a depth of between 55 and 90 cm (Fig. 2.1). Peatlands and soils influenced by groundwater were excluded from the study.

Table 2.1 provides a summary of the characteristics of the sampling sites. The five sandy sites were identified as Cambisols or Podzols and had pleistocene sand as the parent material. The average texture of the reference topsoils was 88$\pm$2% sand, 8$\pm$1% silt and 4$\pm$0.4% clay and the pH was on average 5.3. During the 18th and 19th centuries, the region where the sandy sites are located was characterised by large fen and anthropogenic heath areas (Table 2.2). The sites were completely drained and used for cropping at least a few decades prior to deep ploughing.

The loamy sites were classified as Luvisols, with loess as the parent material. On average, the loamy reference topsoil consisted of 3$\pm$0.4% sand, 82$\pm$1% silt and 15$\pm$1% clay. Mean pH was 6.6. In the region where the loamy sites are located, the land has been used for cropping for several centuries (Table 2.2). Each deep-ploughed subplot and its corresponding reference subplot were uniformly managed by the same farmer after deep ploughing, with the exception of one site (ER) because of changes in ownership. Therefore at site ER, a reference subplot was selected within a directly adjacent field with similar management. At all sites, management was according to common local farming practices (Table 2.2). Crop rotations were dominated by sugar beet and winter wheat at the loamy sites. Crops varied more at the sandy sites, with rye, barley and maize often being grown.

TABLE 2.1: Characteristics of the 10 deep-ploughed cropland sampling sites used in this study

Code	Site[1]	Latitude, Longitude	MAT (°)	MAP (mm)	Elevation (MAMSL[2])	Soil Type[3]	Sand (%)	Silt (%)	Clay[4] (%)	Current ploughing depth (cm)	Deep ploughing depth (cm)	Deep ploughing year[5]
AH	Ahlhorn[a]	52°54'54"N 8°14'59"E	9.0	766	38	Spodic Cambisol[a]	86	10	4	25	90	1968 (46)
BT	Banteln[b]	52°05'13"N 9°44'56"E	9.2	703	86	Haplic Luvisol[b]	5	82	13	32	85	1965 (48)
DB	Drüber[c]	51°45'27"N 9°54'22"E	8.9	687	154	Haplic Luvisol[b]	3	82	15	32	87	1966 (48)
EM	Essemühle[a]	52°45'50"N 8°28'34"E	9.2	717	30	Dystric Cambisol[c]	88	8	4	35	75	1968 (46)
ER	Eickenrode[a]	52°25'57"N 10°20'04"E	9.2	665	54	Gleyic Cambisol[c]	87	8	5	30	65	1968 (46)
EZ	Elze[d]	52°35'06"N 9°45'29"E	9.2	698	38	Dystric Cambisol[c]	84	12	4	27	55	1968 (46)
HB	Hemmelsberg[e]	53°04'59"N 8°19'39"E	9.2	750	10	Haplic Podzol[c]	94	3	3	38	80	1978 (36)
HT	Halchter[f]	52°08'45"N 10°30'31"E	9.0	642	54	Haplic Luvisol[d]	3	83	14	30	70	1966 (48)
SZ	Salzgitter[f]	52°04'12"N 10°27'17"E	9.1	647	116	Haplic Luvisol[b]	3	83	14	33	90	1966 (47)
WB	Warberg[g]	52°11'08"N 10°54'17"E	8.9	649	141	Fragic Luvisol[e]	3	80	17	35	65	1966 (48)

[1] Reference: [a](Foerster, 1974), [b](Bernhard Grosse, 1974), [c](Scheffer and Meyer, 1970), [d](Reichenbach, 1972), [e](Helming, 1987), [f](Czeratzki, 1968), [g](B Grosse and Renger, 1974)
[2] Metres above mean sea level
[3] According to IUSS Working Group WRB (2015), parent material: [a]Pleistocene sand overlying glacial till, [b]Loess, [c]Pleistocene sand, [d]Loess overlying cretaceous lime, [e]Loess overlying limnic clay
[4] Texture values from the reference topsoil
[5] Values in brackets are number of years between deep ploughing and sampling

Soil sampling

Samples were collected to 150 cm depth from one pit each on the deep-ploughed and reference subplots. The pit profile was perpendicular to the deep-ploughing direction. The layers sampled included (Fig. 2.2): (1) the current ploughed horizon; (2) and (3) the deep-ploughed buried topsoil stripes, divided into upper and lower half to assess possible soil depth effects; (4) and (5) the deep-ploughed subsoil stripes, divided into upper and lower half; (6) 10 cm below the deep-ploughed horizon; (7) down to 100 cm; (8) 100 to 120 cm; and (9) 120 to 150 cm.

Representative disturbed samples were extracted for chemical analyses and incubations, and undisturbed samples (three replicates) using horizontal sampling rings of 100 cm^3 volume to determine bulk density. Since results of the chemical characteristics and SOC stability parameters for the upper and lower half of the buried topsoil were mostly identical, we report mean values for both halves of the buried topsoil stripe unless otherwise stated.

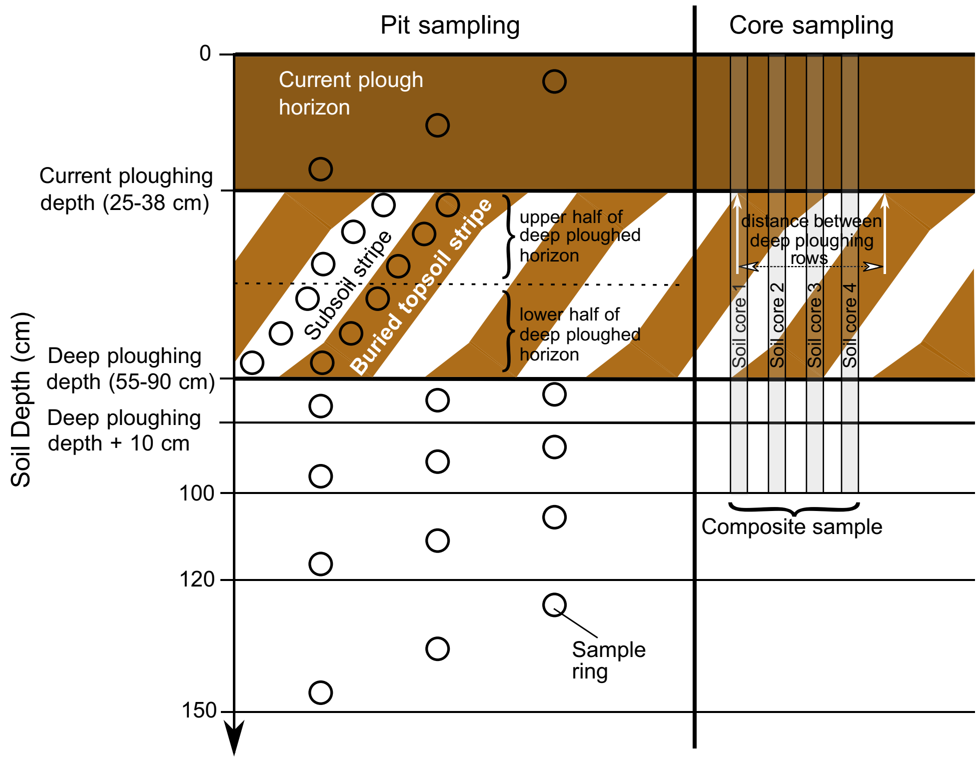

FIGURE 2.2: Pit and core sampling scheme on deep-ploughed subplots. At the pits, 3 sample rings were retrieved from each depth increment down to 150 cm. The deep-ploughed horizon was separated into upper and lower half. Core sampling reached down to 100 cm and the deep ploughing horizon was not divided into two halves. Four cores were distributed along the distance between two deep ploughing rows and constituted one composite sample. At each subplot, 5 composite samples were collected.

In order to account for subplot scale heterogeneity, 20 core samples were collected in the deep-ploughed subplot and 15 in the reference subplot, using a soil auger (60 mm inner diameter; Nordmeyer Geotool GmbH, Berlin, Germany) driven by an electric jackhammer (Wacker EH 23, Wacker Neuson, Munich, Germany). The soil was sampled to 100 cm depth at increments consistent with those ones in pit sampling (1-7), except that the deep-ploughed stripes were not separated into upper and lower halves. If present, compaction and stretching of the cores was recorded and corrected linearly when cutting the soil cores into depth increments. Considering the unequal horizontal distribution of the buried topsoil stripes, we extracted four cores that formed one composite sample at five randomly distributed positions within the deep-ploughed subplot (Fig. 2.2). The position of the four cores was fixed between the deep ploughing rows, so that they captured the inclination of one buried topsoil stripe. The same procedure was followed in the reference subplot, but for only three cores. All sites were sampled between November 2013 and November 2014.

Sample preparation and laboratory analyses

Undisturbed samples from soil rings were dried at 105°C to constant mass and weighed in order to determine soil bulk density. Core and disturbed samples from the profile were dried at 65°C to constant mass and sieved to <2 mm, the fraction used for analysis. Stones and coarse roots removed by sieving were weighed and their volume subtracted from the soil ring volume. Fine roots were removed manually. Aboveground residues were removed directly in the field prior to sampling of soil cores. Loamy samples were coarsely crushed in a jaw crusher (BB1; Retsch, Haan, Germany) instead of sieving. A non-crushed aliquot was used for texture analysis. Soil texture was determined by sedimentation (Moschrefi, 1983) and is given in Table 2.1 for the reference topsoils and Table A3.6 for the buried topsoils.

The pH values were measured in 0.01 M $CaCl_2$ using a glass electrode. For determination of cation exchange capacity, Ca^{2+}, Mg^{2+}, K^+, Na^+ and Al^{3+} were extracted from 100 g soil with 0.1 M $BaCl_2$ (Kretzschmar, 1996), and measured using atomic absorption spectrometry (AAS) (Perkin Elmer AAS 4100, Rodgau, Germany). Dithionite-soluble Fe (Fe_D) was extracted by twice-replicated reduction with $Na_2S_2O_4$ at 85°C (Mehra and M. Jackson, 1958). Amorphous Fe and Al (Fe_O, Al_O) were obtained by extraction with a mixture of ammonium oxalate and oxalic acid in the dark. Concentrations of Fe and Al in the extracts were determined by AAS (AA-280FS, Varian, Palo Alto, CA, USA).

An aliquot of each sieved sample was milled in a planetary ball mill and analysed by dry combustion for total C and total N (TruMac CN LECO, St. Joseph, MI, USA). Samples with a pH value of >6 were analysed for carbonates after ignition of the sample at 450°C for 16 h in a muffle kiln (Bisutti et al., 2007). The SOC content was then calculated by subtracting carbonatic C from total C and expressed as g SOC kg^{-1} soil dry matter.

For the evaluation of SOC stability, we conducted short-term incubation experiments with 3-4 replicates of field fresh samples at 40-50% of their water-holding capacity during 24 h to determine potential mineralisation rate using a semi-automatic continuous flow system (Heinemeyer et al., 1989). Evolving CO_2 was measured hourly with an integrated infra-red analyser (ADC-255-MK3, Analytical Development Co. Ltd., Hoddesdon, England). About 10 to 12 h after starting the incubation, a relatively constant CO_2 respiration rate was reached. Potential SOC mineralisation rate was calculated as the mean of the respiration rates in the second half of the incubation (hours 12 to 14).

Microbial biomass (C_{mic}) was quantified by substrate-induced respiration using 3000 ppm glucose (Anderson and Domsch, 1978) and a conversion factor of 30, as recommended in previous work (E. Kaiser et al., 1992). Potential SOC mineralisation rate is expressed as µg CO_2-C g^{-1} soil dry matter h^{-1}, while C_{mic} values are given as µg C_{mic} g^{-1} soil dry matter. Specific potential SOC mineralisation rate and C_{mic} were obtained through division by the respective SOC content.

Calculations and statistics

Soil bulk density of each depth increment was calculated by dividing the sample masses by the core volume. The mass of roots and stones removed by sieving was subtracted from the sampled soil mass. Plausibility checks were performed with the values obtained for the undisturbed samples collected from the soil profile.

SOC stocks from the soil surface to 100 cm depth were calculated as:

$$\text{SOC stock} = \sum_{i=1}^{n} (\rho_i \times \text{C}_i \times \text{D}_i)$$

where ρ is the bulk density, C is the SOC content and D the size of the depth increment over n depth increments. Considering that deep ploughing involves bulk density changes throughout the soil profile, SOC stocks were corrected for equivalent mass, defining the lightest core of each site as the reference mass (B H Ellert et al., 2007). The values are given as Mg OC ha^{-1}.

To evaluate the overall effect of deep ploughing on SOC stocks, a linear mixed effects model was fitted by restricted maximum likelihood with tillage as a fixed effect with two levels (deep-ploughed vs. reference) and a nested random effect (sites nested within substrates 'loam' and 'sand'). One variance term per tillage substrate combination was applied using package nlme (Pinheiro et al., 2015) in R. The p-value was obtained by conducting an F-Test with the Anova function (Chambers and Hastie, 1992). In the text, the degrees of freedom (df) are also given. For each site, Wilcoxon rank-sum tests were performed to evaluate the effect of deep ploughing on SOC stocks.

SOC stock differences between the reference and deep-ploughed topsoils were also calculated on the basis of equivalent mass. Standard error of the difference between the subplots was calculated following Gauss's law of error propagation (Taylor, 1997). To quantify SOC accumulation in the topsoil of the deep-ploughed subplots, data on SOC contents directly after deep ploughing were taken from the original studies relating to the long-term field trials (Table A3.2). When data were unavailable, the expected SOC content after deep ploughing was approximated as a weighted mean of the SOC contents at different depths before deep ploughing (Table A3.3). It was assumed that there were no bulk density differences after an initial settling phase of the deep-ploughed soil. To assess the long-term SOC dynamics of the buried topsoil after deep ploughing, we obtained initial SOC (before deep ploughing) concentration data from the studies cited in Table 2.1. The SOC contents measured in 2013\14 in the upper and lower buried topsoil stripe were used as the actual concentrations after 36 to 48 years of burial. For site ER, no initial SOC value was available. Decomposition rate constants, assuming C input to be negligible, were obtained using a linear regression fit assuming first-order kinetics (Paul and Clark, 1996).

Differences in potential SOC mineralisation rate, C/N ratio, SOC content, C_{mic} and qCO_2 between the reference and the buried topsoil were assessed by one-way independent samples analysis of variance with the fixed factor horizon (three levels: reference topsoil, upper and lower half of the buried topsoil) and no random factor. The p-values were computed with Tukey's 'Honest Significant Difference' of means on 95% family-wise confidence level with Bonferroni correction (Bretz et al., 2010).

In order to identify factors influencing the effect of deep ploughing on SOC stocks, the same linear mixed effects model mentioned above was used with site as a random effect. The SOC stocks of each core (n=5 in the deep-ploughed and n=5 in the reference subplot) from all sites were set as a dependent variable. Interactions of the tillage effect with the two levels 'reference' and 'deep-ploughed' were assessed as a fixed effect (as in (Walter et al., 2014) with the following variables (properties from the topsoil of the reference subplot; Table A3.4): potential SOC mineralisation rate, specific potential SOC mineralisation rate, C_{mic}, specific C_{mic}, qCO_2, sand, silt and clay content, cation exchange capacity, pH, MAT and MAP (Table 2.1). In addition, the sum Fe_O+Al_O, representing poorly crystalline iron and aluminium oxides and hydroxides, was tested as a variable, as it is related to SOC stabilisation with the formation of organo-mineral complexes (Kleber et al., 2005). The C and N input (kg ha^{-1} yr^{-1}) were also included in this analysis (Organic fertilizer in Table 2.2 converted according to (Döhler, 2009)). A factor was considered to influence the effect of deep ploughing on SOC stocks if it improved the model (verified by Akaike information criterion (AIC) comparison). The significance of the interaction was deduced from the p-values of the model output (Table A3.6). A similar statistical assessment was conducted in order to identify influencing factors on the between-site variability of the specific potential SOC mineralisation rate in the buried topsoil (Table A3.5). All values given in the text are arithmetic means with standard error unless otherwise stated. Differences and effects are significant only when explicitly stated. All statistical analyses were performed with R version 3.2.1 (R Core Team, 2015).

TABLE 2.2: Cropland management at the 10 deep-ploughed cropland sampling sites in the previous 10 years and land use history

Site	Crop rotation[6]	Straw removal	Cover crops	Organic fertilizer application per ha and year[7]	Mineral N fertilization per ha and year[8] (kg)	Subsoil loosening (years; depth)	Land use in the 18th and 19th century[9]
AH	s.B, w.T, w.R	Yes	Turnip, Ryegrass	12 m^3 ps, 1 Mg pm, 17 m^3 d	79	None	Peatland[b,e] (deep-ploughed) heathland[b,e] (reference)
BT	SB, w.W,M	No	Mustard	None	201	None	Cropland[b,e]
DB	RS, w.W, w.B	Partially	None	15 Mg cm, 1 Mg pm, 1Mg tm	163	None	Cropland[b,e]
EM	RS, P, w.B, R,M	Partially	Oilseed radish	16 m^3 cs, 12 m^3 ps, 4 m^3 d, 1 Mg ppl	128	None	Heathland[b,e]
EZ	w.RS, w.R, P	Partially	Mustard	2 m^3 ps, 2 Mg pm	121	None	Heathland[b], cropland[e]
HB	M, w.RS, B, s.B	Yes	Ryegrass	30 m^3 cs, 7 m^3 ps, 2 Mg cm	77	2005; 50 cm	Peatland[c]
HT	SB, w.W, w.B	No	Mustard	1 Mg pm, 8 Mg mm	185	None	Forest[a], cropland[e]
SZ	SB, w.W.	No	Mustard, faba bean	6 m^3 ps, 0.1 Mg sewage sludge	123	2007, 2010; 35 cm	Cropland[d,e]
WB	SB, w.W., w.B.	No	Mustard	2 m^3 ps	182	None	Forest[a], cropland[e]

[6] s.: summer, w.: winter, B: barley, M: maize, R: rye, SB: sugar beet, T: triticale, W: wheat, RS: oilseed rape, P: potato

[7] cs: cattle slurry, cm: cow manure, d: digestate, mm: mixed manure, pm: poultry manure, ppl: potato protein liquid, ps: pig slurry, tm: turkey manure

[8] Values represent the average over 5–11 years according to data provided by farmers

[9] Historical maps: [a] Karte des Landes Braunschweig 1746-1784, [b] Kurhannoversche Landesaufnahme 1773-1783, [c] Oldenburgische Vogteikarte 1791, [d] Gaußsche Landesaufnahme 1827-1840, [e] Preußische Landesaufnahme 1896-1908.

2.4 Results

Long-term effect of deep ploughing on SOM stocks

Deep ploughing significantly increased SOC stocks in the long-term (df=89, F=23, p<0.01). On average, the SOC stocks were 42±18 Mg ha^{-1} higher in the deep-ploughed subplots than in the reference subplots (Fig. 2.3). The SOC stocks in both the deep-ploughed and reference subplots were on average higher at the sandy sites, 210±28 Mg ha^{-1} and 136±11 Mg ha^{-1}, respectively. At the five sandy sites there was high between-site and within-site variability. The loamy sites were more homogeneous in terms of local C distribution, with average SOC stocks in the deep-ploughed subplots of 78±2 Mg ha^{-1} and in the reference subplots of 67±3 Mg ha^{-1}. The greatest difference (289±48 Mg ha^{-1}) was observed at site AH, while the smallest non-significant difference (7±17 Mg ha^{-1}) occurred at site HB (Table A3.1). Lower SOC stocks (3±4 Mg ha^{-1}) in the deep-ploughed subplot compared to the reference subplot were only found at site DB.

Nitrogen was sequestered together with SOC, with the deep-ploughed subplots having on average 1.8±0.6 Mg ha^{-1} higher stocks than the reference subplots (Fig. 2.3). At the sandy sites the N stocks in the deep-ploughed subplots were 2.8±1.0 Mg ha^{-1} higher than in the reference subplots, while at the loamy sites the difference amounted to 0.8±0.4 Mg ha^{-1}.

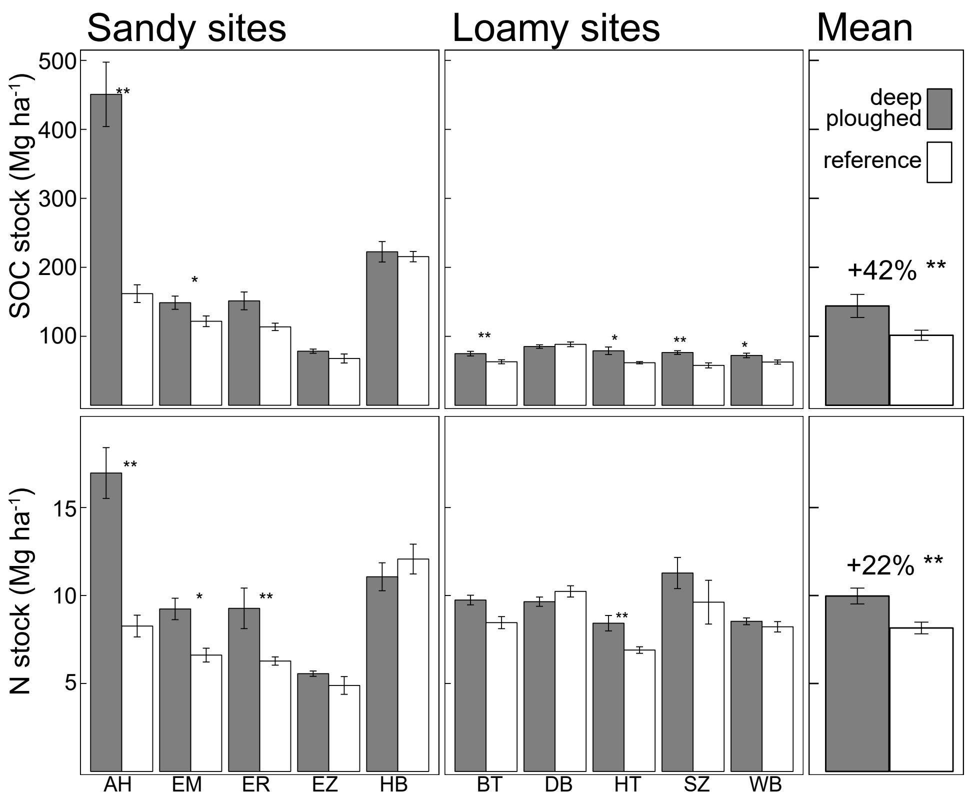

FIGURE 2.3: SOC and N stocks down to 100 cm in deep-ploughed and reference subplots at the sandy and loamy sites as determined from core sampling. Bars represent standard errors of the mean (n = 5 at each subplot and site). Mean values of all sites represent the relative difference between all cores from deep-ploughed (n = 50) and reference subplots (n = 50). Significance levels according to Wilcoxon rank sum tests: **P <0.01, *P <0.05

The contribution of the subsoil (below the current plough horizon) to total SOC stocks in the reference soils was 23±3% at the sandy sites and 27±4% at the loamy sites. In the deep-ploughed subplots, subsoil SOC contributed 45±8% and 38±4% to the total SOC stocks (0-100 cm) for sandy and loamy sites, respectively, indicating increased importance of the subsoil for SOC storage.

Accumulation of SOC in topsoil after deep ploughing

Through deep ploughing, SOM-poor subsoil material is mixed into the topsoil, which initially decreases topsoil SOC stocks, forming a SOC accumulation potential in this 'new' topsoil. After 45 years on average, the SOC stocks in the topsoil of the deep-ploughed subplots were 4 to 22% lower than in the reference topsoils. After the deep ploughing event, the new topsoil of the deep-ploughed fields accumulated on average 0.4±0.1 Mg SOC ha^{-1} yr^{-1} (Table A3.3). There were no significant differences in the SOC accumulation rates between loamy and sandy topsoils.

The SOC stocks in the loamy deep-ploughed topsoils were 4±1 Mg ha^{-1} lower than in the reference topsoils (Fig.2.4). At the sandy sites the deficit was greater, with 17±4 Mg ha^{-1} lower SOC stocks in the deep-ploughed topsoils. The largest deficit was found at site HB (33±15 Mg ha^{-1}), which is the site with the shortest period of time since the deep ploughing was performed.

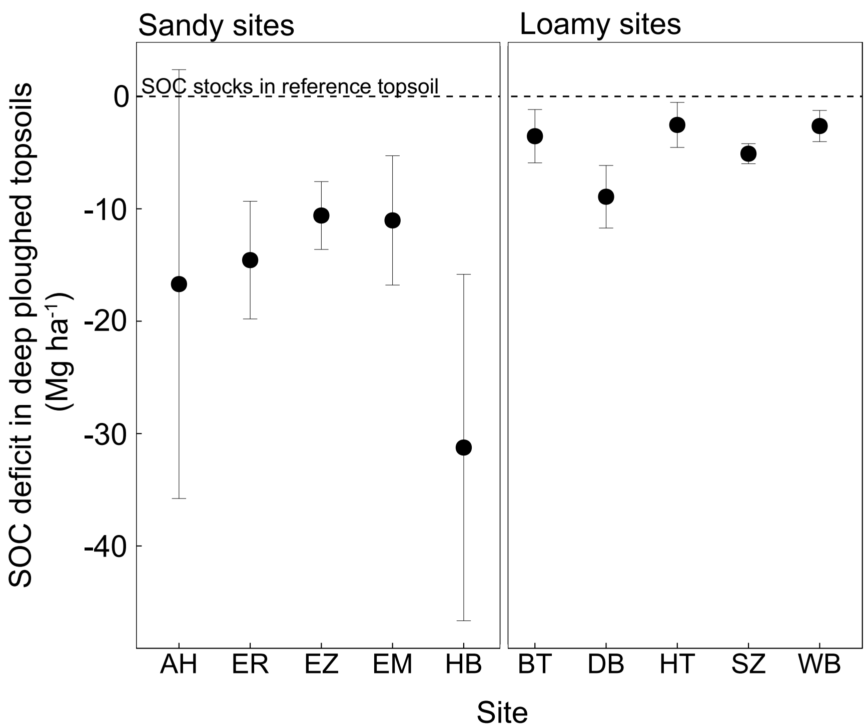

FIGURE 2.4: SOC stock deficit in deep-ploughed topsoil at the sandy and loamy sites compared with the SOC stocks in the reference topsoil.

SOC stability in buried and reference topsoils

Deep ploughing led to the burial of SOM-rich topsoil material down to 90 cm depth. The SOC contents observed in this study provided an insight into the stability of the buried SOC (Fig. 2.5). The buried topsoil stripes (27±5 g kg^{-1}) at the sandy sites did not contain significantly less SOC than the reference topsoil (25±5 g kg^{-1}). In contrast, SOC contents in the buried topsoil stripes (6±0.2 g kg^{-1}) at the loamy sites were significantly lower than in the reference topsoil (10±0.7 g kg^{-1}). However, the SOC contents were 156% higher in the loamy buried topsoil stripes than in the subsoil horizons at the same depth in the reference soil.

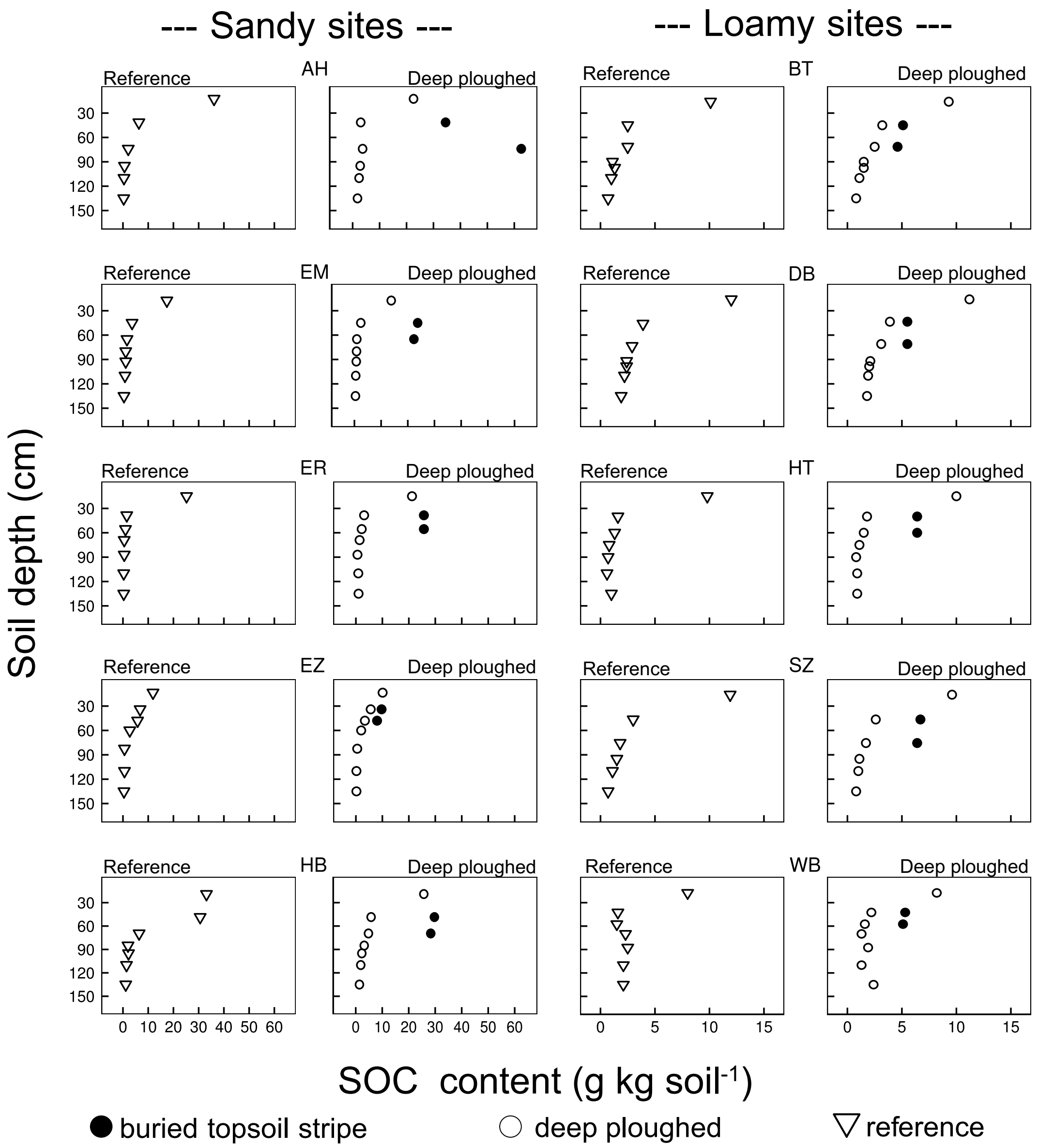

FIGURE 2.5: Depth distribution of SOC contents in deep-ploughed and reference subplot pits for each sandy and loamy sampling site. Buried topsoil stripes in the deep-ploughed horizon were sampled with soil rings separately from the surrounding subsoil (see Fig. 2.2). To assess possible soil depth effects, one sample was collected from the upper half and one from the lower half of the buried topsoil stripes.

The stability of the buried SOC was assessed by comparing the initial SOC content (non-buried topsoil) with the current SOC content in the buried topsoil stripes using historical reports (Fig. 2.6). On average 45 years after deep ploughing, 47 to 73% of the initial SOC still remained in the buried topsoil stripes, assuming no significant SOC input into the subsoil. At sites AH and EM, we even recorded higher SOC contents in the buried topsoil than in the topsoil before deep ploughing, indicating uncertainty in the historical data, but also very small or no changes in the SOC contents.

At the loamy sites, the SOC content in the buried topsoil decreased by 0.07 to 0.13 g kg^{-1} yr^{-1} and between 47 and 60% of the initial SOC was maintained due to burial. The SOC content of the buried topsoil at the sandy site EZ was also within this range. At site HB the SOC content decreased faster (by 0.31 g kg^{-1} yr^{-1}), with 73% of the initial SOC being maintained 36 years after topsoil burial due to deep ploughing. At the five loamy sites and at sites EZ and HB, the SOC contents in the upper half of the buried topsoil (9.8 g kg^{-1}) were slightly higher than in the lower half (9.2 g kg^{-1}), as a result of lower decomposition rates in the upper half (0.12 g kg^{-1} yr^{-1}) than in the lower half (0.14 g kg^{-1} yr^{-1}) of the buried topsoil stripe. These differences were not statistically significant, however.

FIGURE 2.6: Long-term change in SOC content since topsoil burial by deep ploughing for seven sampling sites compared with bare fallow field experiments. Mean and standard error of the mean (n = 7) are shown. Mean and standard error of bare fallow (n = 7) calculated with data from [1] (Barré et al., 2010).

Potential SOC mineralisation assessed in short-term in vitro incubations supported the findings on the long-term SOC dynamics. The potential SOC mineralisation rate was significantly lower in the buried sandy topsoil (0.11±0.02 µg CO_2-C g^{-1} h^{-1}) than in the non-buried reference topsoil (0.33±0.06 µg CO_2-C g^{-1} h^{-1}) (Fig. 2.7a). In contrast, the mineralisation of buried SOC (0.21±0.06 µg CO_2-C g^{-1} h^{-1}) at the loamy sites was only slightly and non-significantly lower than in the reference topsoil (0.29±0.10 µg CO_2-C g^{-1} h^{-1}).

The differences between sandy and loamy soils and between the buried and reference topsoils were even larger considering the potential mineralisation per amount of SOC (Fig. 2.7b). The specific potential SOC mineralisation rate was significantly lower in the buried sandy topsoils (5±1 µg CO_2-C g^{-1} SOC h^{-1}) than in the reference topsoils (16±5 µg CO_2-C g^{-1} SOC h^{-1}). However, in the buried loamy topsoil the specific potential SOC mineralisation rate was higher (36±9 µg CO_2-C g^{-1} SOC h^{-1}) than in the reference topsoil (27±3 µg CO_2-C g^{-1} SOC h^{-1}).

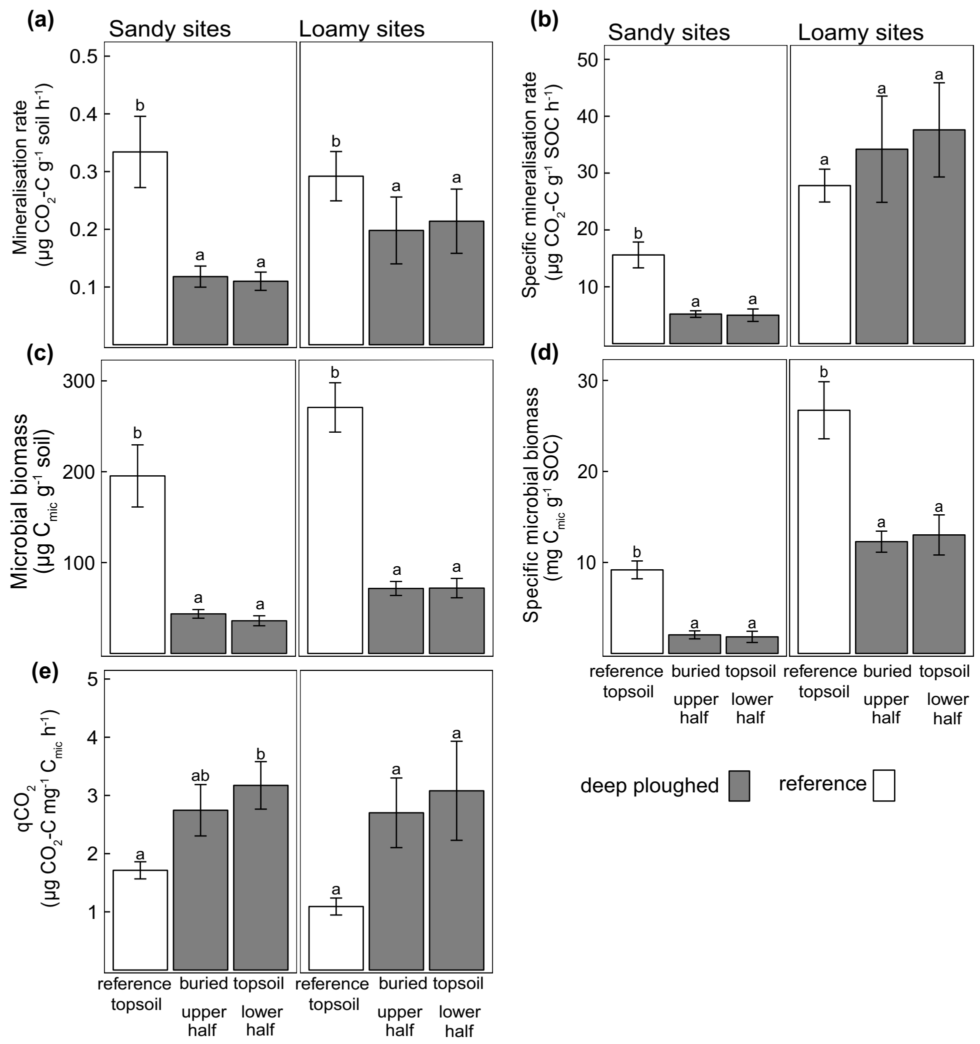

FIGURE 2.7: Potential SOC mineralisation and microbial biomass in the reference and buried topsoils at the sandy and loamy sites. **(a)** Potential SOC mineralisation rate, **(b)** specific potential SOC mineralisation rate, **(c)** $_{mic}$, **(d)** specific C_{mic} and **(e)** qCO_2 of buried and reference topsoils of loamy and sandy soils. Buried topsoil was separated into upper and lower halves. Bars represent standard errors of the mean, calculated separately from the mean values of sandy (n = 5) and loamy sites (n = 5). Different letters represent significant differences (P <0.05) within the sandy and loamy soils, respectively.

The microbial biomass was significantly reduced through topsoil burial. At the sandy sites, C_{mic} in the buried topsoil was 40±5 µg g^{-1}, compared with 195±34 µg g^{-1} in the reference topsoil (Fig. 2.7c). However, C_{mic} was even lower in the adjacent subsoil stripes (21±6 µg g^{-1}) and in the reference subsoil (17±3 µg g^{-1}). A similar pattern was observed at the loamy sites, but with higher levels of C_{mic} (72±8 µg g^{-1} in the buried topsoil stripe compared with 271±27 µg g^{-1} in the reference topsoil). Similarly, lower values were found in the subsoil stripes (53±5µg g^{-1}) and in the reference subsoil (38±2 µg g^{-1}).

Normalising C_{mic} per amount of SOC as specific C_{mic} indicates the amount of microbes living per unit SOC (Fig. 2.7d). The specific C_{mic} in the buried topsoil (2±0.5 mg g^{-1} SOC) at the sandy sites was significantly lower than in the reference topsoil (9±1 mg g^{-1} SOC). Similarly, the specific C_{mic} in the buried topsoil (13±2 mg g^{-1} SOC) at the loamy sites was significantly lower than in the reference topsoil (26±3 mg g^{-1} SOC).

The qCO_2 values, i.e. the potential SOC mineralisation per unit microbial biomass (Fig. 2.7e), revealed that SOC mineralisation in the buried topsoil at the sandy sites was significantly more efficient (2.9±0.4 µg CO_2-C mg^{-1} C_{mic} h^{-1}) than in the reference topsoil (1.7±0.2 µg CO_2-C mg^{-1} C_{mic} h^{-1}). At the loamy sites, there was no significant difference between buried and non-buried topsoil but a similar trend emerged, since qCO_2 in the buried topsoil was higher (2.9±0.7 µg CO_2-C mg^{-1} C_{mic} h^{-1}) than in the reference topsoil (1.1±0.2 µg CO_2-C mg^{-1} C_{mic} h^{-1}).

The C/N ratio served as an additional indicator of SOC mineralisation dynamics. Although no significant differences in C/N were found between the buried and reference topsoils (Fig. 2.8), the C/N ratio differed markedly between the sandy and loamy soils. At the sandy sites, buried topsoil (24±3) had a wider C/N ratio than the reference topsoil (17±5). In contrast, at the loamy sites, the C/N ratio narrower in buried topsoil (8±0.3) than in the reference topsoil (10±0.5).

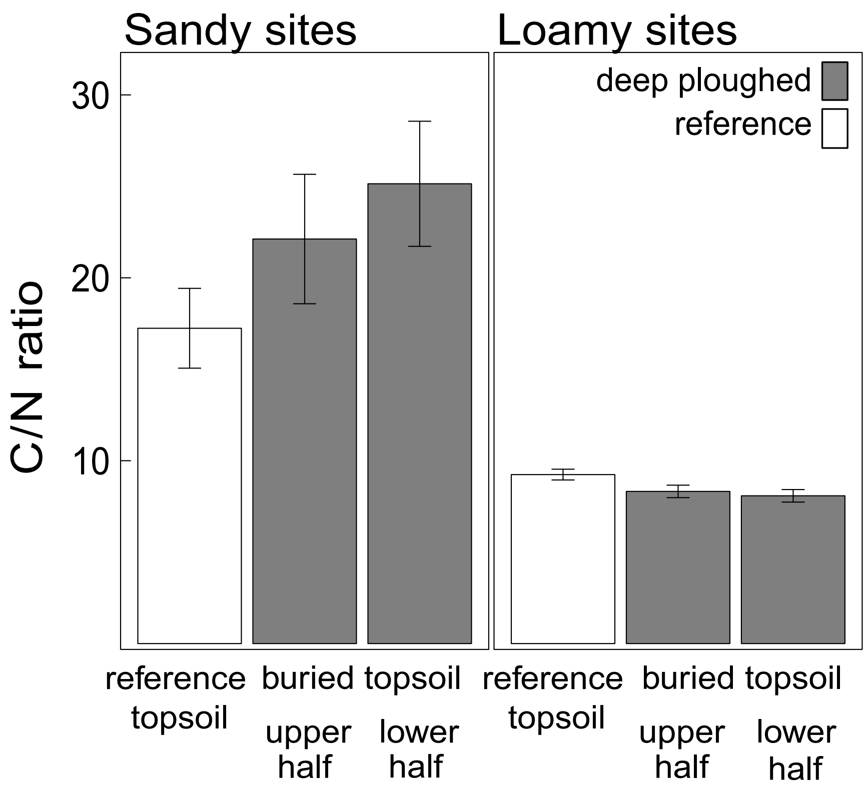

FIGURE 2.8: C/N ratio of reference and buried topsoils. Buried top- soil was separated into upper and lower halves. Bars represent standard errors of the mean, calculated separately from the mean values of sandy (n = 5) and loamy sites (n = 5). There were no significant differences (P <0.05) within the sandy and loamy soils.

Factors influencing the effect of deep ploughing on SOC stocks

The influence of site characteristics (climate and management variables) and soil properties on the effect of deep ploughing on SOC stocks was assessed by a mixed effects linear model. All factors tested increased the explanatory power of the variance of the positive effect of deep ploughing on SOC stocks, except for N fertilisation and absolute and relative microbial biomass. However, only four of these factors significantly explained the variability between the 10 sites. The enhancing effect of deep ploughing on SOC stocks was significantly lowered (negative effect) by increasing Fe_{D-O}, silt and clay content (Table A3.5) and positively influenced by the sand content. Thus, the main differences in SOC accumulation after deep ploughing at the 10 sites were due to the different texture and the content of crystalline iron oxides. The content of crystalline oxides was particularly high at site DB (4.7 g kg^{-1}), which was the only site with lower SOC stocks in the deep-ploughed subplot than in the reference subplot.

Furthermore, explanatory variables for the specific potential SOC mineralisation rate of the buried topsoil in the deep-ploughed subplots were assessed using the same factors. Similarly to the change in SOC stocks, mineralisation rate of the buried topsoil decreased with increasing sand content (Table A3.7), emphasising the higher stability of SOC at the sandy sites. In addition, higher pH and higher silt and clay content were significantly related to the higher specific SOC mineralisation rate that characterised the loamy sites.

2.5 Discussion

Increased SOC stocks due to deep ploughing

SOC-rich topsoil material buried in the subsoil was not completely decomposed 36 to 48 years after deep ploughing, resulting in 3 to 178% higher SOC stocks (42 Mg ha^{-1} on average) in the first 100 cm depth (Fig. 2.3). Deep ploughing proved to be a measure to enhance SOC sequestration, by translocating large amounts of not easily decomposable SOC to greater soil depth. This has been recommended as a means of tapping the SOC sequestration potential of subsoils (Lorenz and Lal, 2005; C. Rumpel et al., 2012).

Based on our results, the SOC sequestration due to deep ploughing is more than 10-fold the effect of conversion from conventional to continuous no-tillage practices. A previous study showed that with no-tillage, on average only 3.4 Mg SOC ha^{-1} were accumulated in the first 30 cm of the soil within 15 years (Virto et al., 2012). The average SOC sequestration rate within 45 years since deep ploughing (0.96 Mg ha^{-1} yr^{-1}) was two- to four-fold higher than the estimated SOC storage effect of no-till, which is between 0.23 and 0.4 Mg ha^{-1} yr^{-1} (Freibauer et al., 2004).

Elevated SOC stocks in buried soils with high SOC contents have also been reported for volcanic soils with former topsoil buried under new volcanic sediments (Basile-Doelsch et al., 2005; Peña-Ramírez et al., 2009; Zech et al., 2014). Buried topsoil in depositional sites of eroding landscapes have also been characterised as C sinks (K Van Oost et al., 2007). Deposition areas with buried topsoil material following erosion have been found to contain higher SOC stocks than the soil at the eroded site (VandenBygaart et al., 2012).

The highest SOC stocks in the deep-ploughed subplots (Fig. 2.3) and the highest measured SOC content in the buried topsoil stripes (Fig. 2.5) were found at site AH. Although the soil at this site is described as a Dystric Cambisol (Foerster, 1974), historical landscape maps from the 18th century designate the site as a fen area (Table 2.2). According to the farmer, the fen was drained and cultivated decades before deep ploughing of the field. This could explain the unexpectedly high SOC stock at this site. Moreover, it shows that deep ploughing has helped to preserve the relict SOC of the fen, since the SOC content in the buried topsoil stripes was even higher than in the reference topsoil. The fact that the lowest specific C mineralisation rate among all samples was found for the buried topsoil stripes at site AH and that the differences in the magnitude of this parameter compared with in the reference topsoil were greatest at this site also reflect the resistance of the buried SOC to decomposition.

Site HB stored only 3% more SOC in the deep-ploughed subplot than in the reference subplot. The reference subplot contained the highest SOC stocks of all 10 sites studied. HB was the only site at which deep subsoiling to 50 cm depth was conducted (in 2005; see Table 2.2) and was the site with the highest application rate of liquid manure (cattle and pig slurry). A positive relationship between the amount of animal manure applied and SOC stocks has been reported (Maillard and D. A. Angers, 2014), which could explain the relatively high SOC stocks at HB. The subsoiling in 2005 also led to SOC translocation to 50 cm depth in both the reference and deep-ploughed subplots. However, since the SOC stocks in the deep-ploughed topsoil were relatively depleted (Fig. 2.4) compared with the reference topsoil, enhancement of the SOC stocks in the subsoil to 50 cm depth due to subsoiling took place presumably only in the reference subplot. An indicator of this is the high SOC content in the reference subsoil (Fig. 2.5), which can also have derived from the existing Bh-Horizon. This explains the relatively small difference in SOC stocks between the deep-ploughed and reference subplots compared with the other study sites.

Site DB was the only site at which deep ploughing did not enhance SOC stocks (Fig. 2.3). The SOC content in the buried topsoil was similar to that in the reference subsoil, which was higher than that at all other loamy sites (Fig. 2.5). Similarly to site HB, the reference subplot at site DB contained the highest SOC stocks of all loamy sites (Fig. 2.3), both in topsoil and subsoil. The C input with organic fertilisers at site DB was the highest of all five loamy sites (Table 2.2). This could explain the high SOC stocks in the reference subplot, considering the positive relationship between animal manure application and SOC stocks (Maillard and D. A. Angers, 2014). This is also reflected in the relatively high SOC accumulation rate of the deep-ploughed topsoil (0.36 Mg ha^{-1} yr^{-1}; Table A3.3). Since the SOC stocks were high not only in the topsoil but also in the subsoil of the reference subplot, it can be assumed that downward transport SOC is relatively high at site DB. Beside natural processes and conditions, high SOC stocks can be caused by temporarily deeper regular ploughing (e.g. to 40 cm) followed by shallower ploughing (at DB currently 32 cm; Table 2.1). This could in turn promote mineralisation of the buried SOC in the deep-ploughed subplot, which would explain why site DB had the greatest change in buried topsoil SOC content (0.13 g kg^{-1}soil yr^{-1}; Fig. 2.6). Nevertheless, this site was also that with the highest SOC accumulation potential in the topsoil, since it had the highest SOC deficit in the topsoil of the deep-ploughed subplot among the loamy soils (Fig. 2.4).

SOC accumulation in the new topsoil after deep ploughing

Besides translocation of SOC-rich topsoil material into the subsoil (below the recently ploughed layer), the SOC sequestration effect of deep ploughing can be attributed to long-term SOC accumulation in the newly formed SOC-poor plough layer (Fig. 2.1). The new plough horizon with an initially reduced SOC content provided additional storage space for SOC. Carbon-free mineral surfaces in the subsoil are the components responsible for accumulation and stabilisation of SOC (Rasse et al., 2005). For sandy soils, we assumed that a significant proportion of the SOC was accumulated as particulate organic matter, since its proportion relative to SOC commonly increases with increasing sand content (Liang et al., 2003).

The effect of incorporation of subsoil material into the newly formed plough layer by deep ploughing can be compared with the effect of deepening of the plough horizon from 25 to 35 cm, which was promoted in many countries in Europe during the 1960s and 1970s. About three decades after deepening of the topsoil, SOC contents present before onset of deeper ploughing were almost re-established and SOC stocks reached higher levels than the initial SOC stocks (Nieder and Richter, 2000). About 10-20 Mg SOC ha^{-1} accumulated in deepened plough layers of different farming systems. In 'flipped' soils in New Zealand, which were inverted with an excavator down to 1 to 3 m depth, topsoil SOC contents increased nearly five-fold from, around 7.5 g kg^{-1} to 35 g kg^{-1}, within 8 years after the flipping (Thomas et al., 2007). In another study in New Zealand, the SOC content in fresh slip scars (eroded surface areas) in hill country increased from 2 g kg^{-1} to 3-4 g kg^{-1} within 20-30 years (Rosser and Ross, 2011).

In the present study, significant differences in topsoil SOC stocks between the non-deep-ploughed reference and deep-ploughed subplots persisted even on average 45 years after the deep ploughing event (Fig. 2.4). Thus our hypothesis that the SOC content present before deep ploughing would be re-established in the newly formed topsoil within 40 to 50 years after deep ploughing could not be confirmed. Slow SOC accumulation in SOC-depleted new topsoil is in line with observations of SOC accumulation and losses after land use changes. SOC can be lost quickly after land use changes but SOC accumulation, e.g. after afforestation, is slow, more than 100 years, following the pattern of 'slow in, fast out' (Poeplau et al., 2011). On eroded uplands with subsoil exposure at the soil surface, SOC accumulation (Berhe et al., 2007) may also be slower than expected, given the persistent difference in topsoil SOC observed here between deep-ploughed soil and the reference soil.

The SOC dynamics of arable soils are related directly to the C input through roots, residue return and organic fertilisation. Management, e.g. organic fertilisation, was the same for deep-ploughed and reference subplots in this study, since they were both part of one uniformly managed field. Differences in C input between the two subplots may have arisen from different crop yields and thus different inputs through roots and residue return. However, no consistent effect on crop yield due to deep ploughing at the studied sites AH, BT, DB and EM was reported (Scheffer and Meyer, 1970; Foerster, 1974; Bernhard Grosse, 1974). The limited long-term effects on yield on most soils and for most crop types was the main reason why deep ploughing has not been continued on a larger scale since the 1970s. When a long-term increase in yield was reported, it was due to enhanced infiltration through subsoil loosening and thus reduced flooding damage to crops (Baumhardt et al., 2008). However, the loosening effect of deep ploughing in Luvisols persisted for only one year (Hartge, 1981). A long-term decrease in yield on deep-ploughed croplands has not been reported.

The N stocks in the deep-ploughed subplots were on average 22% higher than in the reference subplots. Thus, the relative N sequestration was lower than the SOC sequestration. The additional N needed to increase the SOC stocks after deep ploughing (Fig. 2.3) amounted to around 41 kg N ha^{-1} yr^{-1} or 1.8 Mg N ha^{-1} in total.

Buried SOC stability at sandy and loamy sites

SOC in the buried topsoil at the sandy sites was found to be more stable than in the reference topsoil, since the SOC content was higher (Fig. 2.5), the potential mineralisation rate was lower (Fig. 2.7a,b) and the C/N ratio were wider (Fig. 2.8). Contrasting results were obtained on the loamy soils, where the buried topsoil stripes contained less SOC (Fig. 2.5), had higher potential mineralisation rate (Fig. 2.7a,b) and narrower C/N ratio (Fig. 2.8) than the reference topsoil. This was unexpected, since SOC stabilisation has often been found to be related particularly to fine-textured soils (Flessa et al., 2008; Doetterl et al., 2015a).

Lower specific SOC mineralisation rates have been found in sandy historical heath soils with partly high groundwater levels in north-west Germany than in reference Cambisols (Springob and Kirchmann, 2002). It has been shown that heath litter and SOC under heath vegetation contain high concentrations of alkyl C, mainly comprising lipids, waxes, resins and suberin, and have relatively high hydrophobicity index (Certini et al., 2015), impeding microbial decomposition of the organic material (Bachmann et al., 2008). Cropland soils on historical heath still contain large proportions of lipids, although they may be not as high as in permanent heath (Sleutel et al., 2008). Inhibited SOC decomposition has been attributed to the presence of phytotoxic and fungitoxic substances in Calluna heath soils (Jalal and Read, 1983).

For heath soils, wide C/N ratio of >30 (Rowe et al., 2006) or >20 (Certini et al., 2015) has been observed. Wide C/N ratio is also common in peat soils (Aitkenhead and McDowell, 2000), but in cultivated peatlands C/N ratio can decrease as a consequence of drainage and agricultural management (Krüger et al., 2015). Values of C/N ratio found for the sandy soils in our study (24±3) were in the range reported in the latter study (Fig. 2.8). Nitrogen limitation can also explain the low SOC mineralisation rate (Trinsoutrot et al., 2000). On the other hand, the high N fertilisation rate (Table 2.2) does not indicate that N availability was limited. Furthermore, it has been shown that the SOC mineralisation rate is lower in soils with lower pH (between 4.5 and 8.3) due to higher fungal-to-bacterial biomass ratio (Rousk et al., 2009). The sandy sites in the present study were located in a region with a high abundance of peatland (3500 km^2) in the 18th century, of which 80% was under agricultural use by 1960 (Overbeck, 1975). Anthropogenic heath area in northern Germany was greatest at the end of the 18th century and has been drastically reduced since the beginning of the 19th century due to land reforms and cultivation with deep ploughing parenciteBehre2008. Using historical maps, we were able to confirm a history of peatland or heath for all five sandy sites (Table 2.2), which may explain the high stability of the SOC. The land use history conditioned the build-up of chemically recalcitrant and hydrophobic SOC, which we concluded was the major factor for stabilising the SOC in buried sandy topsoil.

Deep ploughing might also change other soil parameters such as the hydrological regime, with higher water storage capacity in the subsoil and increased rooting depth in the buried topsoil stripes, which could also provide nutrients. However, during sampling we did not observe different rooting patterns in the deep-ploughed soils compared with the reference soils, since we mainly sampled during winter.

There was no evidence that SOC in the buried topsoil at the loamy sites was more stable than that in the reference topsoil. However, except for site DB, the buried topsoil stripes at these sites also had higher SOC contents than the adjacent subsoil stripes and the subsoil of the reference subplot at equivalent depth (Fig. 2.5). Bioturbation through earthworms may play a role in SOC translocation in loamy soils. However, we observed very few earthworm burrows at the loamy sites and none at the sandy sites. Even at grassland sites with much more burrows, the earthworm effect on subsoil SOC stocks has been found to be small (Don et al., 2008). Due to the smaller differences between the topsoil of the deep-ploughed and reference subplots in the loamy compared with the sandy soils (Fig. 2.4), it can be inferred that in the newly formed topsoil of the deep-ploughed subplots, a larger amount of SOC was accumulated. This explains why the deep-ploughed subplots at four out of five loamy sites studied had higher SOC stocks than the reference subplots (Fig. 2.3). In summary, we can conclude that deep ploughing leads to long-term SOC storage in loamy soils too.

Comparison of the SOC contents in the topsoil before burial through deep ploughing and in the buried topsoil stripes on average 45 years after deep ploughing (Fig. 2.6) indicated that only part of the original SOC was mineralised (40-53% in loamy soils and 28-39% in sandy soils). The buried topsoil stripes were decoupled from new C input, similarly to bare fallow. In previous long-term bare fallow experiments, the average SOC decomposition rate was 0.16±0.02 g kg^{-1} yr^{-1} in the first 20-25 cm depth within an average of 46 years (Barré et al., 2010), which is higher than the estimated decomposition rate of SOC in the buried topsoil stripes of 0.13±0.03 g kg^{-1} yr^{-1} in our study. This emphasises the stabilising effect of burial on SOC. On excluding site HB, for which we assumed higher C input into the subsoil (Table 2.2), mean SOC decomposition rate in the buried topsoil at the remaining sites was on average 0.10±0.01 g kg^{-1} yr^{-1}. This is significantly lower than the SOC decomposition rate in bare fallow topsoil. At two sites (AH and EM), a decrease in the SOC content in the buried topsoil could not be confirmed, leading us to conclude that the buried SOC at those sites is markedly stable over the long-term. Enhanced root litter input cannot explain the high SOC contents in the buried topsoil stripes, since the low mineralisation rate in these stripes indicates that the SOC present is stabilised and old, and not fresh SOC input from roots or dissolved organic carbon.

With our hypothesis that the buried SOC was more stable than that in the reference topsoil, we expected that Cmic would be lower in the buried topsoil stripes. Overall, this was the case, even regarding the amount of C_{mic} per unit SOC (specific C_{mic}; Fig. 2.7d). However, this did not mean lower SOC mineralisation rate, since the qCO_2 values were higher in the buried than in the reference topsoil. This is in line with patterns reported in other studies for subsoil Cmic and microbial efficiency (Lavahun et al., 1996; Jörgensen et al., 2002).

Factors influencing the effect of deep ploughing on SOC stocks

Assessment of the influence of soil properties and site characteristics showed that the effect of deep ploughing on SOC stocks was significantly lowered with increasing content of crystalline iron and with increasing silt and clay content (Table A3.5). Soil texture and the content of crystalline iron are closely related to each other, as reflected by the fact that the increase in SOC stock through deep ploughing was lower in the loamy soils than in the sandy soils. Interestingly, the crystalline iron content was on average five-fold higher in the loamy than in the sandy reference topsoil (Table A3.4). Similar trends in crystalline iron content have been reported for Luvisols, Arenosols and Podzols in Poland (Różański et al., 2013). Iron oxides are mainly part of the clay-sized soil fraction (McFadden and Hendricks, 1985; Eusterhues et al., 2003) and thus their effect on SOC stabilisation cannot be separated from the effect of the clay content. On the other hand, the sand content positively influenced the deep ploughing effect on SOC stocks (Table A3.5), underscoring that in sandy soils enhancement of SOC stocks, especially through higher storage in the subsoil, is limited and may be facilitated by deep ploughing much more than on loamy soils. The mineralisation rates determined in incubation experiments were also closely related to soil texture, with the lowest values in sandy soils (Tables S5 and S6). Moreover, the pH was significantly positively related to the mineralisation rate of the buried topsoil SOC, which is in line with previous findings (Rousk et al., 2009). Thus, the stability of SOC in sandy soils was inherent and deep ploughing did not reduce it.

All other variables considered had an influence, although not significant, on the enhancement of SOC stocks through deep ploughing, except for the microbial biomass. This is a further indication that C_{mic} content per se is not directly related to SOC mineralisation.

Deep ploughing contributes to SOM sequestration

Subsoils have been proven to have great potential for SOM storage in many studies and their importance within the global C cycle has been highlighted. Our study showed that deep ploughing can contribute to additional SOM sequestration, with higher SOC stocks in the long-term (on average 42 Mg ha^{-1}). Deep ploughing increased the storage space for SOM by mechanically translocating topsoil material into the subsoil, where originally relatively less SOM is stored in most soil types. The contribution of the subsoil to total SOC stocks to 100 cm depth was 11-22% greater in deep-ploughed soils than in reference soils. Thus the translocation of SOM into the subsoil not only increased subsoil SOC stocks, but also full profile SOC stocks. This was due to the establishment of new topsoil horizons that under constant management continued to accumulate SOC, even after more than four decades. In contrast to most measures for sequestering SOC on managed soils, deep ploughing needs to be conducted only once to induce long-term SOC accumulation. Even after four decades, the topsoil SOC contents in the deep-ploughed subplots were still lower than in the reference subplots, indicating that these newly formed topsoils will presumably continue accumulating SOC. Surprisingly, the effect of deep ploughing on C stocks and SOC stability after burial was more pronounced at the sandy sites than at the loamy sites. During the history of the sandy sites as heath, hydrophobic SOC was formed, which remained very stable after topsoil burial. Thus, these sandy sites may not be directly comparable to other sandy sites with a different land use history. Conveniently, at sandy sites deep ploughing is less energy-intensive and more beneficial for soil structure.

The energy consumption during deep ploughing operations has to be taken into account when considering deep ploughing as a climate change mitigation option. An initial decrease in soil fertility through SOM dilution in the newly formed topsoil horizon is to be expected. Moreover, deep ploughing is generally not possible in soils which are not deeply developed. However, subsoils provide a largely unused source for nutrients that becomes available for plants after deep ploughing. Furthermore, deep ploughing may enhance the rooting depth and water storage capacity and thus stabilise crop yields under climate change conditions.

3 Stability of buried carbon in deep ploughed forest and cropland soils - implications for carbon stocks

3.1 Abstract

Accumulation of soil organic carbon (SOC) may play a key role in climate change mitigation and adaptation. In particular, subsoil provides a great potential for additional SOC storage due to the low SOC content and assumed higher stability of subsoil SOC. However, the underlying drivers for higher subsoil SOC stability remain unclear. The fastest way in which SOC reaches the subsoil is via burial to greater soil depth, e.g. via erosion or in cropland via deep ploughing. In this study, we assessed the effect of active SOC burial through deep ploughing on long-term SOC stocks and stability in forest and cropland subsoil at 12 sites. After 25-48 years, deep-ploughed subsoil contained significantly more SOC than reference subsoils, in both forest soil (64±9 and 43±6 Mg ha^{-1}, respectively) and cropland (40±3 and 24±3 Mg ha^{-1}, respectively). However, total SOC stocks down to 100 cm in deep-ploughed soil were greater than in reference soil only in cropland, and not in forests. This was mainly explained by slower SOC accumulation in topsoil of deep-ploughed forest soils compared with cropland. Buried SOC was on average 32% more stable than reference SOC, as revealed by long-term incubation. Moreover, buried subsoil OC had higher apparent radiocarbon ages indicating that it is largely isolated from exchange with atmospheric CO_2. We concluded that deep ploughing can increase SOC storage in subsoils and that the higher SOC stability in subsoil is not only a result of SOC ageing combined with selective preservation of more stable SOC fractions.

3.2 Introduction

Soil organic carbon (SOC) is currently receiving increasing attention in science and politics due to its great potential to act as a sink for atmospheric CO_2 and thus mitigate climate change (Paustian et al., 2016). SOC may also help to adapt to climate change because of its beneficial effect on soil structure, water-holding capacity and nutrient retention (Dalal et al., 2011). Apart from land use conversions such as afforestation, SOC accrual can be achieved through implementation of certain management practices including conservation agriculture, cover crop cultivation and mulch farming, among many others (Freibauer et al., 2004; Lal et al., 2015). Currently, most SOC sequestration management measures are based on assessment of the increase in SOC content in the top layers of the soil. However, over half of world's total SOC is located below 30 cm depth, in the subsoil (Batjes, 2014; Hiederer and Köchy, 2012). Although SOC concentration decreases with depth, SOC stocks in subsoil are mostly greater than in topsoil because subsoil has a larger soil mass, and thus larger potential storage capacity, than topsoil.

Subsoil OC is reported to be more stable than SOC near the soil surface, a trend inferred from increasing apparent radiocarbon age with depth, indicating that deep SOC has prevailed in soils and has been excluded from exchange with the atmosphere for centuries to millennia (Gleixner, 2013; Mathieu et al., 2015). It has been widely suggested that subsoil has great potential to store additional SOC than topsoil (Lorenz and Lal, 2005; C. Rumpel et al., 2012) because of the large number of unsaturated mineral surfaces (Beare et al., 2014) and environmental conditions that slow SOC mineralisation (Rumpel et al., 2011) (e.g. more constant moisture and temperature regime or oxygen limitation). Additional carbon inputs have been observed to decompose more slowly in subsoil than in topsoil (Wordell-Dietrich et al., 2016). These slower SOC mineralisation rates in subsoil have been attributed to the lower SOC content, which results in a lower density of decomposing microorganisms and thus a lower possibility of any SOC present being mineralised (Don et al., 2013).

Carbon enters subsoil mainly with aboveground and belowground litter, dead roots and root exudates, dissolved and particulate organic carbon (OC) transported via large pores or through biological soil reworking (bioturbation) (C. Rumpel et al., 2012). However, deeper burial of C-rich soil material, e.g. by deposition following erosion, generally leads to a long-term increase in landscape-scale SOC stocks (Doetterl et al., 2016). Active anthropogenic SOC burial has rarely been studied, but is commonly carried out through deep ploughing of agricultural and forest land.

Deep ploughing is a land management operation performed only once, with the purpose of loosening the subsoil, enhancing water infiltration and root penetration capacity and thus improving plant growing conditions. As a preparation measure for afforestation, deep ploughing leads to a higher survival rate of planted trees because of better weed control, with weed seeds being buried, and better water availability when the OC-rich A horizon with high water-holding capacity is placed deeper in the soil (Hansen et al., 2007). Through the action of deep ploughing, SOC-rich topsoil is buried at 60-120 cm depth and SOC-poor subsoil material is brought up to the surface. The latter also transports nutrients from the less weathered subsoil to the surface, making them more easily available to plants. Deep ploughing of cropland is reported to be a very effective long-term SOC sequestration measure (Alcántara et al., 2016). At 10 cropland sites on mineral soils, deep ploughing led to a 42% increase in SOC stocks after 45 years because carbon in the buried topsoil was not entirely mineralised, if at all, and additional SOC was continuously accumulated in the "newly formed" topsoil mixed with subsoil material.

Because deep ploughing translocates large amounts of SOC to the subsoil and also facilitates deep rooting, subsoil SOC stocks can be expected to increase over the long-term, including in forests. Deep ploughing of forest soil also leads to burial of the organic layer that forms on top of the mineral soil – the forest floor - with its additional carbon. Rooting patterns are also different in forest compared with cropland with deeper and more roots in forest soil (R. B. Jackson et al., 1996). In the present study, we investigated the effect of carbon burial through deep ploughing in forest and cropland soils. In addition, because harvesting is four-fold more intense for crops than for forests (Schulze et al., 2010), SOC in the newly formed topsoil of deep-ploughed soil can be expected to accumulate faster in forest soil than in cropland. On the other hand, fertilisation, liming and tillage of cropland may stimulate SOC accumulation in topsoil of the deep-ploughed arable soils.

The following hypotheses were tested in this study:

1. SOC stocks increase on a long-term basis after deep ploughing compared with non-deep-ploughed reference soil. This SOC accrual is greater in forests than in cropland.
2. Buried SOC is more stable to mineralisation than non-buried SOC in reference topsoils.

3.3 Methods

Study sites and sampling

Twelve experimental sites were selected for sampling, namely four forest sites and three cropland sites on sandy soils and five cropland sites on loamy soils (Table 3.1). Each site comprised a deep-ploughed plot and an adjacent reference, non-deep-ploughed plot with plot size 20m×40 m. All other site factors, such as forest and cropland management, soil characteristics,tree species and crops were equal or very similar in both plots (Table 3.1). Rebberlah was the only site that was not an experimental field site, but was partially deep-ploughed after a wildfire. The Lindenburg (Schrey and Bergfeld, 1985) and Schwenow sites were clear-cut and partially deep-ploughed for experimental purposes regarding soil loosening and thus improvement of tree growth conditions. The Viborg site, located in Jutland, Denmark, is part of an experimental site studying different site preparation measures for afforestation of former arable land. The other three sites were located in northern and eastern Germany (Table A4.1).

A joint comparative analysis with five loamy and three sandy deep-ploughed cropland sites presented in Alcántara et al. (2016) was performed. Soil sampling was conducted in a similar way as in that study by taking five soil cores in each plot down to 100 cm to assess SOC stocks. These cores were divided into four depth increments: (1) Forest topsoil with high SOC content, (2) subsoil down to deep ploughing depth (Table 3.1), consisting of alternating buried topsoil stripes and subsoil stripes in the deep-ploughed plots (3) deep-ploughing depth + 10 cm and (4) deep subsoil down to 100 cm. Forest floor was sampled prior to coring with 25cm×25cm metal frames separating horizons into L (undecomposed leaf litter) and F+H (partly decomposed organic matter). After drying at 65°C, sieved roots and stones were weighed to obtain root biomass and fine soil mass. In addition, undisturbed and disturbed soil samples for chemical analyses were taken from soil profiles. This enabled separated sampling of the buried topsoil stripes.

Chemical and microbiological analyses

Basic chemical and microbial characterisation was conducted as for the deep-ploughed cropland sites used for comparison (Alcántara et al., 2016). SOC mineralisation was assessed in a one year batch incubation at 22°C of 100 g soil dry matter in triplicate gas-tight 250 mL glass flasks. Water content was initially adjusted to 60% of the water-holding capacity of each soil and gravimetrically readjusted periodically to the initial water content. CO_2 production was measured at day 1, 3, 8, 14, 31, 127, 195, 269 and 365 after incubation start. CO_2 concentration in sampled 20 mL vials was measured with a gas chromatograph (Series GC-2014; Shimadzu Deutschland GmbH, Duisburg, Germany). Specific CO_2 production per g SOC as a cumulative sum of the total incubation year was computed.

TABLE 3.1: Characteristics of deep ploughed forest and cropland sites. Soil properties refer to the reference topsoil (texture: n=1, pH: n=6). Soil types according to IUSS Working Group WRB (2015) d.p.: deep ploughing, r.:reference.

Land use	Site	D.p. year	Years since d.p.	D.p. depth (cm)	Dominant tree or crops	Soil type	Sand (%)	Silt (%)	Clay (%)	pH	Former land use
Forest	Lindenburg	1977	37	60	*Pinus sylvestris*, *Quercus robur*, *Fagus sylvatica*	Spodic Cambisol	86	9	5	4.5±0.3	Forest
Forest	Rebberlah	1978	36	58	*Pinus sylvestris*, *Picea abies*	Lamellic Podzol	84	11	5	3.1±0.1	Heathland\ Forest
Forest	Schwenow	1961	53	60	Quercus rubra (d.p.), Pinus sylvestris (r.)	Haplic Podzol	89	11	0	4.4±0.7	Heathland
Forest	Viborg	1989	25	62	Quercus robur, Pinus sylvestris	Haplic Cambisol	86	9	5	4.8±0.2	Heathland, Cropland
Cropland	Elze	1968	46	55	Oilseed rape, rye, potato	Dystric Cambisol	84	12	4	5.3±0.1	Cropland
Cropland	Essemühle	1968	46	75	Oilseed rape, potato, barley, rye, maize	Dystric Cambisol	88	8	4	4.8±0.1	Heathland
Cropland	Hemmelsberg	1978	36	80	Oilseed rape, rye, potato	Dystric Cambisol	94	3	3	5.3±0.05	Peatland
Cropland	Banteln	1965	48	85	Sugar beet, wheat, maize	Haplic Luvisol	5	82	13	6.6±0.1	Cropland
Cropland	Drüber	1966	48	87	Oilseed rape, wheat, barley	Haplic Luvisol	3	82	15	6.6±0.04	Cropland
Cropland	Halchter	1966	48	70	Sugar beet, wheat, barley	Haplic Luvisol	3	83	14	6.5±0.1	Cropland
Cropland	Salzgitter	1966	47	90	[l] Sugar beet, wheat	Haplic Luvisol	3	83	14	6.9±0.03	Cropland
Cropland	Warberg	1966	48	65	Sugar beet, wheat, barley	Fragic Luvisol	3	80	17	6.0±0.1	Cropland

Density fractionation (Cerli et al., 2012; Golchin et al., 1994) of SOC was performed by suspending 30 g soil dry matter in 120 mL of 1.6 g cm^{-3} sodium polytungstate (SPT) to separate a free light fraction (fLF). Dispersion by ultrasound with an energy input of 400 J ml^{-1} (Schmidt et al., 1999) and resuspension in SPT was applied to obtain an occluded light fraction (oLF). A heavy fraction (HF) remained as sediment. Radiocarbon content in oLF and HF was assessed by acceleration mass spectrometry with preceding sample preparation (Rethemeyer et al., 2013) and calibrated to Fraction Modern (Reimer et al., 2004).

Calculations and statistics

Calculed SOC stocks (Poeplau and Don, 2013) $\left[\text{Mg ha}^{-1}\right]$ were corrected for different masses to enable comparison on an equivalent mass basis (B. H. Ellert and Bettany, 1995). Data analysis to identify significant differences between deep-ploughed and reference soils was performed with R (R Core Team, 2015) version 3.3.1. Normality of data was first checked with the Shapiro-Wilk test. If data sets were normally distributed, differences were evaluated with paired Student's t-tests. Otherwise, Wilcoxon Rank Sum and Signed Rank tests were applied. When repeated observations per site were made, i.e. sampling of five cores per plot in each site, linear mixed effect models using package nlme (Pinheiro et al., 2016) were computed with plot (reference and deep-ploughed) as fixed effect and site as random effect. If necessary, variances were weighed to ensure homoscedasticity. Correlations between stability indicators were examined with Spearman correlation tests. Fraction Modern values of ^{14}C were converted to absolute Fraction Modern (Trumbore et al., 2016) and then to $\Delta^{14}C$ using R package SoilR version 1.1-23 (Sierra et al., 2014).

3.4 Results

Depth distribution of SOC contents and stocks

Buried topsoil stripes in deep-ploughed forest soil had higher SOC contents (16±4 g kg^{-1} soil) than the corresponding depth layer in adjacent subsoil stripes (4±1 g kg^{-1} soil) and the reference subsoil (7±2 g kg^{-1} soil) (Fig. 3.1). This was also observed for the cropland soils studied (16±5, 3±0.4 and 4±2 g kg^{-1} soil in deep-ploughed, adjacent and reference subsoil, respectively). At two sites, SOC content in the buried topsoil was comparable to that in the reference topsoil: for the Schwenow forest site these values were 29±10 and 26±4 g kg^{-1} soil, respectively, and for the Essemühle cropland site they were 17±6 and 23±1 g kg^{-1} soil, respectively. At all other sites, SOC content in the buried topsoil stripes was reduced by 11% (Hemmelsberg) to 95% (Lindenburg) in comparison with the non-buried reference topsoil. Assuming that, before deep ploughing, the SOC content in the buried topsoil stripes was similar to that in the current reference topsoil, the fastest decrease in SOC content in buried topsoil stripes following deep ploughing was at the Lindenburg and Rebberlah forest sites (mean annual decrease of 2.6% and 2%, respectively).

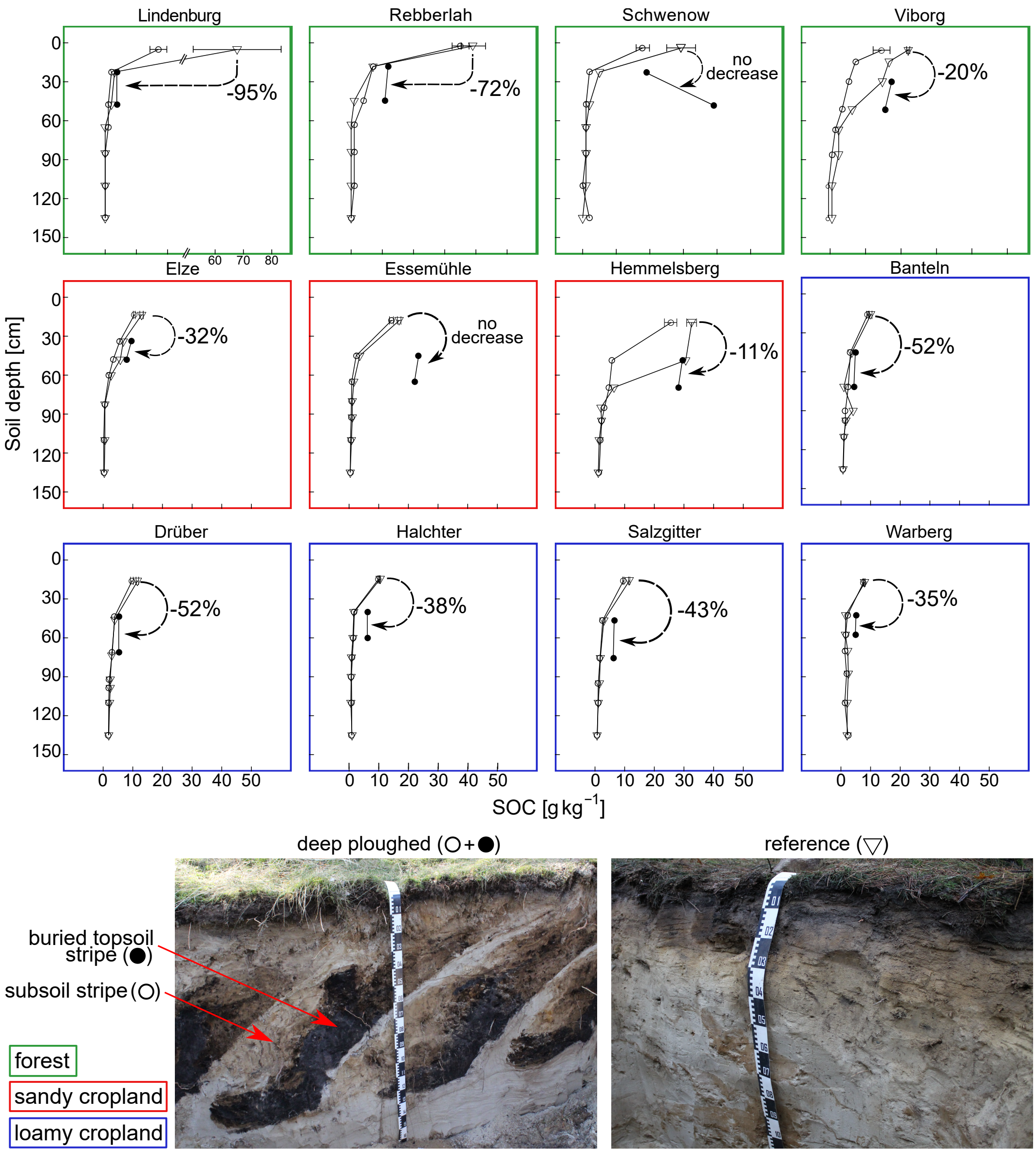

FIGURE 3.1: Depth distribution of mean SOC content in deep-ploughed (○ + ●) and reference plots (▽). Topsoil SOC content in soil profiles and cores (N=6). Deep-ploughed plots consist of alternating buried topsoil (●) and subsoil stripes (○). Dashed arrows and percentages indicate the relative difference between average topsoil SOC and average SOC in buried topsoil stripes.

Newly formed topsoil on deep-ploughed cropland soil had 8% lower SOC stocks than the reference topsoil (57±5 and 64±6 Mg ha^{-1}, respectively, p=0.02). An even larger difference of 37% was observed in forest soils (20±1 and 32±3 Mg ha^{-1}, respectively, p<0.0001). Forest floor followed the same trend, but differences were only significant in the F+H-horizons (6±1 Mg ha^{-1} in deep-ploughed and 15±3 Mg ha^{-1} in reference plots, p=0.03). On average, the difference between topsoil SOC stocks in deep-ploughed and reference soils relative to the number of years since deep ploughing was -0.16 Mg ha^{-1} yr^{-1} in cropland and -0.31 Mg ha^{-1} yr^{-1} in forest soil. Moreover, the nitrogen (N) stocks in topsoil of deep-ploughed soil were substantially lower than in reference soil in forests, while at five out of eight cropland sites studied, the topsoil of deep-ploughed soils contained more N than reference topsoil (Table A4.1).

Total SOC stocks down to 100 cm were significantly greater in deep-ploughed than in reference cropland soil (Fig. 3.2, 105±8 and 92±8 Mg ha^{-1}, respectively, p<0.0001). Below 30 cm down to deep-ploughing depth, deep-ploughed cropland subsoil contained 67±17% more SOC than reference subsoil (40±3 and 24±3 Mg ha^{-1}, respectively, p<0.0001). In contrast, total SOC stocks in deep-ploughed forest soil (including forest floor) were not significantly greater than in reference soil (103±11 and 105±9 Mg ha^{-1}, respectively, p=0.2). However, SOC stocks in forest subsoil were 49±25% greater in deep-ploughed than in reference soil (64±9 and 43±6 Mg ha^{-1}, respectively, p=0.0002). The SOC stocks below the deep-ploughed horizon did not differ between treatments within forest or cropland sites.

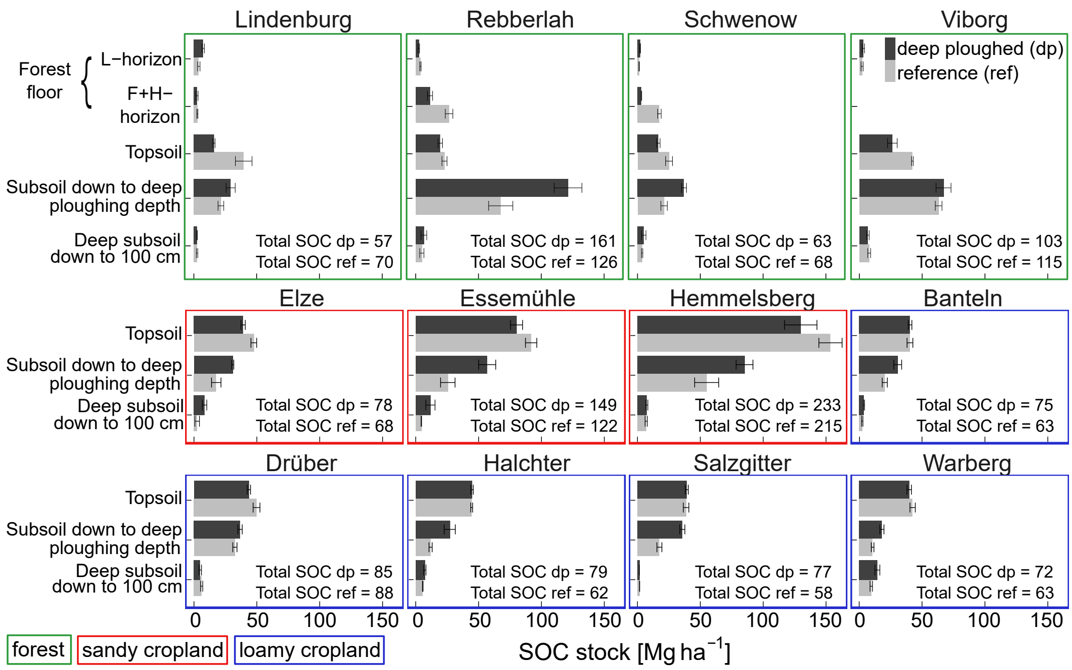

FIGURE 3.2: SOC stocks at different soil depth increments in deep-ploughed and reference plots. Bars represent average SOC stocks in soil cores (n=5), whiskers show standard error. Subsoil and buried topsoil stripes were not sampled separately. Total SOC stock sums for forest sites include forest floor.

Potential SOC mineralisation

Buried SOC stability was assessed through one-year incubation experiments, which enabled comparison of SOC turnover in buried topsoil stripes and reference topsoil under standardised laboratory conditions eliminating possible oxygen or water limitations. The fraction of mineralised SOC was 32% lower in incubated buried topsoil than in reference topsoils (Fig.3.3, p≈0). Forest soils had the highest mineralisable SOC fraction, both in buried topsoil stripes and in reference topsoil (56±13 and 77±22 mg CO_2-C g^{-1} SOC, respectively). Sandy cropland buried topsoil stripes and reference topsoil had the lowest mineralisable SOC fraction (27±4 and 40±6 mg CO_2-C g^{-1} SOC, respectively). There was a weak positive correlation between the relative difference in specific cumulative SOC mineralisation and the relative difference in SOC content for the buried topsoil stripes and the reference topsoil (Fig. 3.6a, Rho = -0.6, p = 0.04).

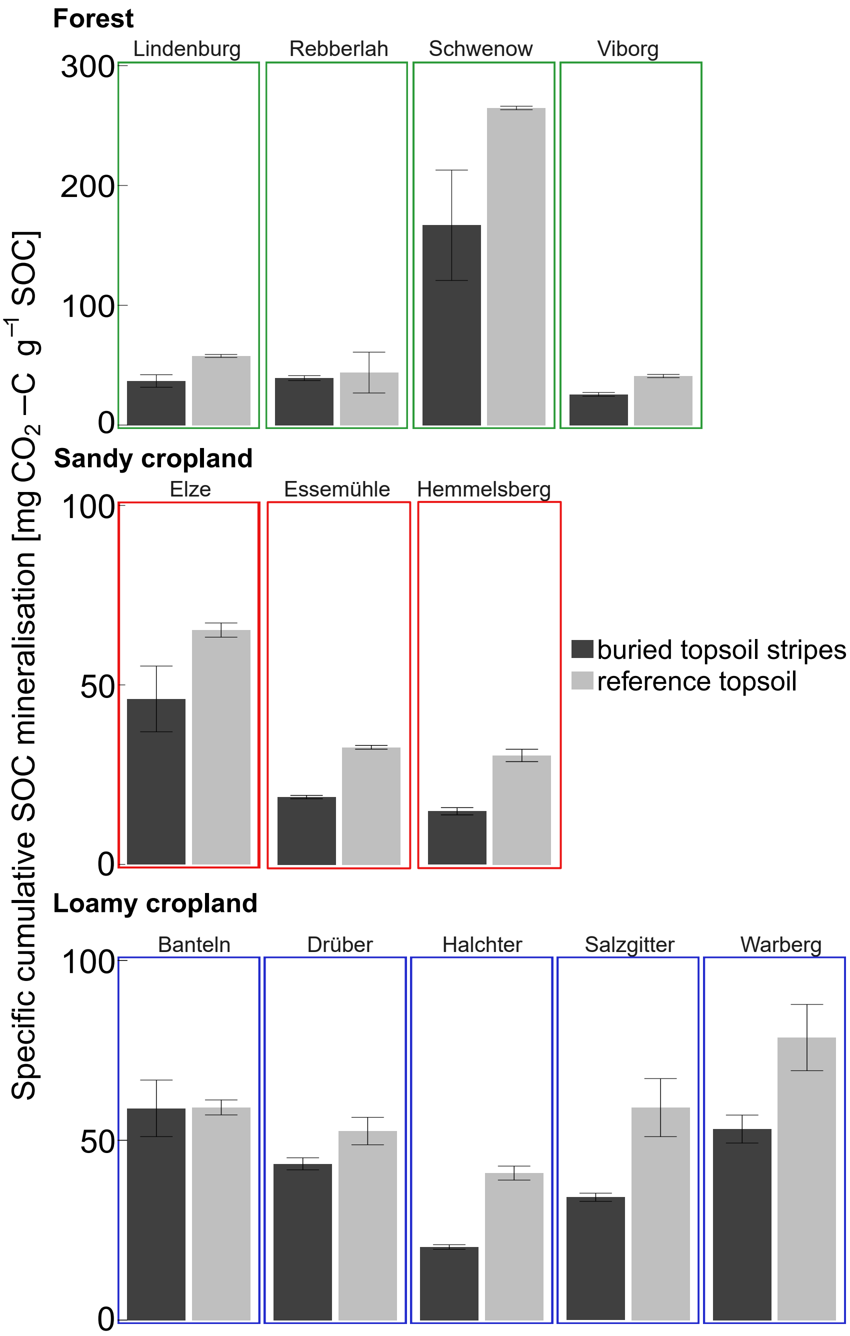

FIGURE 3.3: Specific cumulative SOC mineralisation after one year of incubation. Bars represent mean values from laboratory replicates (n=3), whiskers show standard error.

Carbon input and SOC fractions

The relative distribution of different SOC fractions provided information about the degree of stabilisation of carbon in the buried topsoil stripes and the reference topsoil. The free light fraction (fLF) is usually the youngest and most labile SOC fraction. However, the SOC content in fLF was not consistently lower in buried topsoil stripes than in reference topsoil, but rather ranged between 65% lower and 19% higher (Fig. 3.4). The correlation between the effect of burial on specific SOC mineralisation and the effect of burial on fLF mass was very weak and not significant (Fig. 3.6b, Rho = -0.3, p = 0.3). Similarly, SOC in occluded light fraction (oLF) was between 90% lower and 70% higher in buried topsoil stripes than in the reference topsoil. The most stable heavy fraction (HF) did not consistently contain most of the SOC in buried topsoil. Instead, SOC in the HF was between 28% lower and 79% higher in buried topsoil compared with reference topsoil.

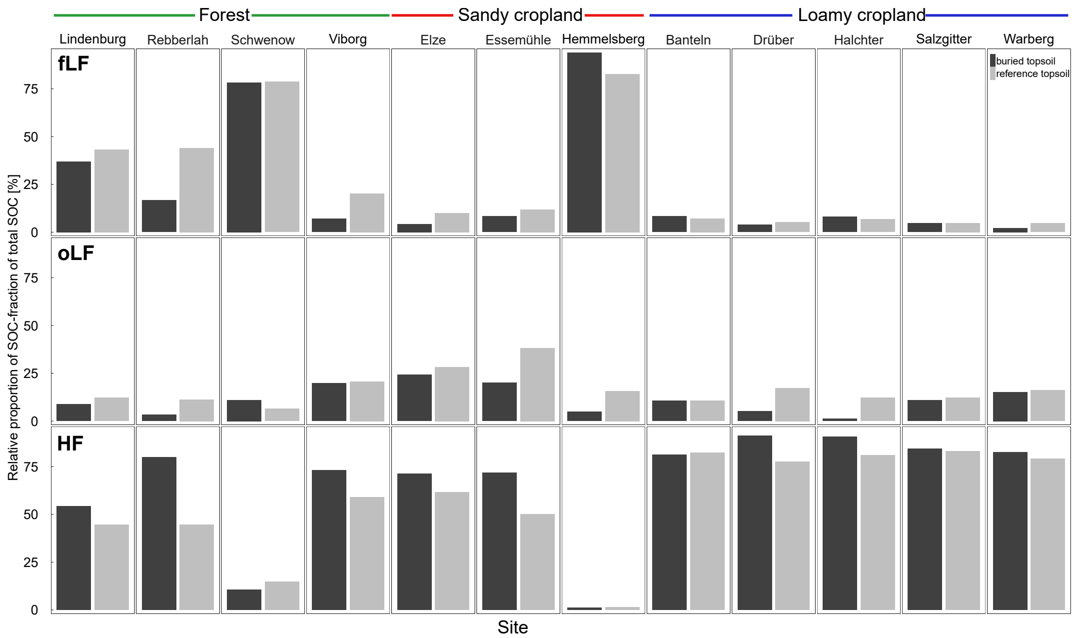

FIGURE 3.4: Relative proportion of free light fraction (fLF), occluded light fraction (oLF) and heavy fraction (HF) of total SOC in buried topsoil stripes and reference topsoil.

Fine root biomass was used as an indicator for carbon input as labile SOC. In general, root biomass was 10 to 100 times lower in cropland than in forest soil (Fig. 3.5a). Throughout the entire soil profile, root biomass in cropland did not differ between deep-ploughed and reference plots. In contrast, root biomass in forest topsoils tended to be lower in deep-ploughed plots than in reference topsoil (13±3 and 15±3 g kg $^{-1}$, respectively). Deep-ploughed forest subsoil had 65% higher root biomass than reference subsoils (1.5±0.3 and 0.9±0.11 g kg $^{-1}$, respectively, p=0.04). At the Viborg site only, relative root mass was very similar in subsoil of both plots. Root biomass in the deep-ploughed subsoil was highly correlated to the fLF content (Fig. 3.5b).

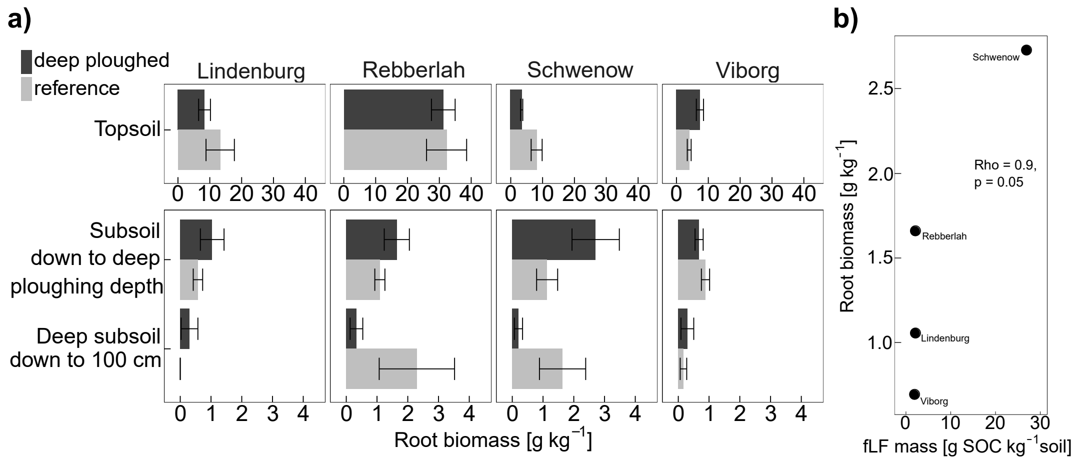

FIGURE 3.5: **a)** Root biomass at different soil depth increments in deep-ploughed and reference forest soil. Bars represent average root biomass in soil cores (n=5), whiskers show standard error. **b)** Correlation between root biomass in deep-ploughed subsoil and fLF mass calculated as Spearman's rank correlation.

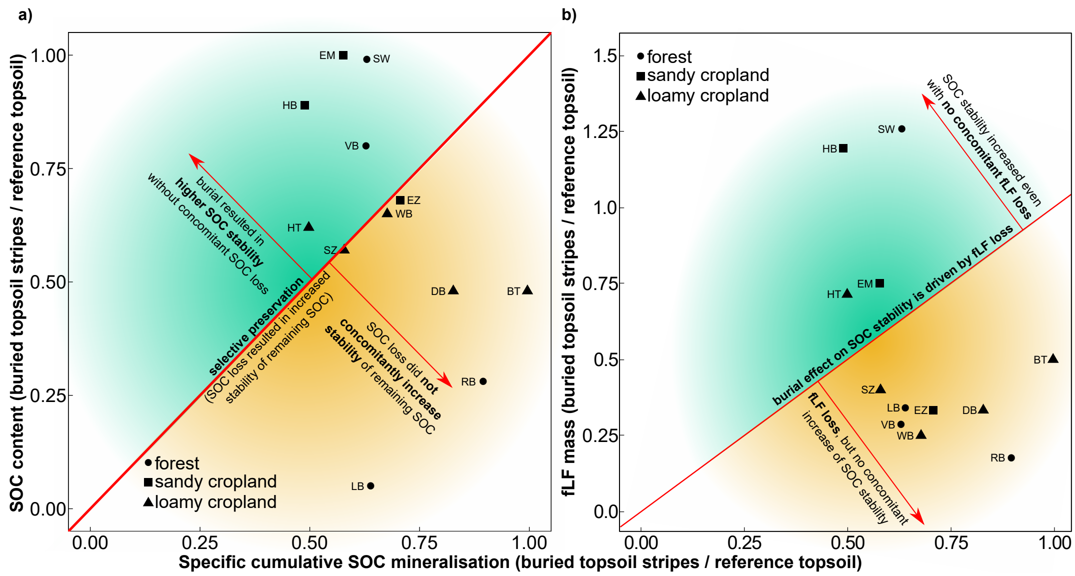

FIGURE 3.6: Correlation between ratio of specific cumulative SOC mineralisation in buried topsoil stripes to that in reference topsoil. **a)** ratio of SOC content in buried topsoil stripes to that in reference topsoil as well as **b)** ratio of fLF mass in buried topsoil stripes to that in reference topsoil. Sites abbreviations: LB - Lindenburg, RB - Rebberlah, SW - Schwenow, VB - Viborg, EZ - Elze, EM - Essemühle, BT - Banteln, DB - Drüber, HT - Halchter, SZ - Salzgitter and WB - Warberg

Radiocarbon content

The radiocarbon (^{14}C) content of the SOC fractions oLF and HF provided information about the mean residence time of carbon in the soil. Longer residence times are generally characterised by a low ^{14}C content. The oLF had higher ^{14}C content than HF indicating faster turnover of the oLF (-71 and -76‰, respectively). The oLF of the buried topsoil stripes had lower $\Delta^{14}C$ values than that of reference topsoil at all sites except the Viborg forest site and the Halchter loamy cropland site (Fig. 3.7, -120±23 and -22±15‰, respectively). The ^{14}C content in the oLF was lower in buried topsoil stripes than in reference topsoil, by 96‰in forest, 45‰in sandy cropland and 133‰in loamy cropland. This pattern of lower ^{14}C content in buried SOC compared with reference topsoil was also observed for HF (-117±14 and -34±18‰, respectively). The ^{14}C content in HF was lower in buried topsoil stripes than in reference topsoil by 74‰in forest, 30‰in sandy cropland and 121‰in loamy cropland. In both fractions, as well as in buried topsoil stripes and reference topsoil, $\Delta^{14}C$ was higher in forests than in cropland (Fig. 3.7). Forested reference topsoil had mostly positive $\Delta^{14}C$ values, indicating the influence of deposition of ^{14}C deriving from nuclear weapons testing since the 1960s (oLF: -6±15, HF: 12±16‰).

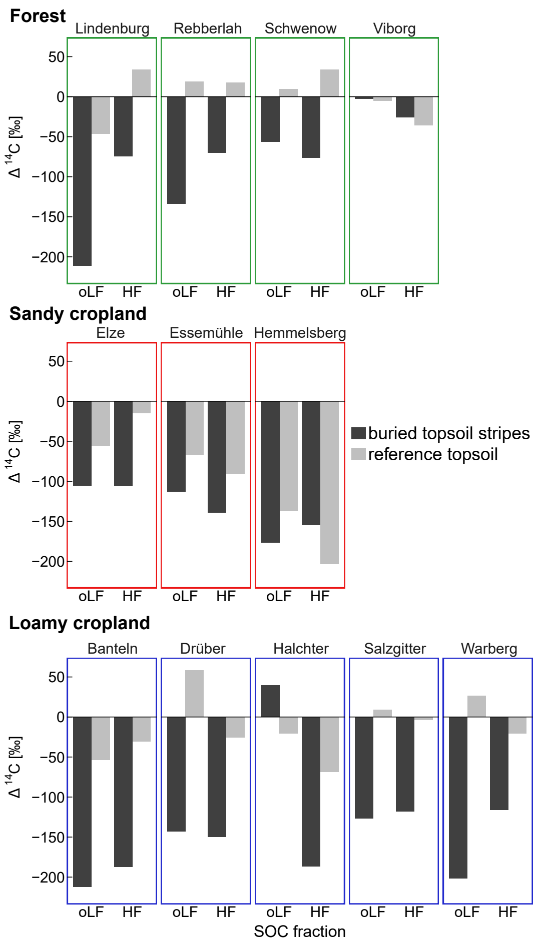

FIGURE 3.7: $\Delta^{14}C$ values obtained for occluded light fraction (oLF) and heavy fraction (HF) of SOC in buried topsoil stripes and reference topsoil. Positive values signify that nuclear test- derived ^{14}C was incorporated into the SOC fraction (young apparent ^{14}C age). Negative values indicate that soil fraction carbon exchange with the atmosphere has been slow and that significant radioactive decay has occurred (old apparent ^{14}C age) (Trumbore et al., 2016).

3.5 Discussion

The SOC stocks in subsoil were significantly greater in deep-ploughed soil than in reference plots (Fig. 3.2). The data also revealed that SOC was partly preserved, in both forest and cropland soils, since topsoil burial up to 53 years ago (Table 3.1). In colluvial deposition areas under cropland, topsoil buried to 30-70 cm depth has been reported to contain more SOC 50 years after burial than the corresponding topsoil in upslope areas (VandenBygaart et al., 2012). In these environments, SOC mineralisation presumably takes place over a timescale of centuries, with 50% of the buried SOC mineralised after ca. 250–300 years (Wang et al., 2014). However, a certain fraction (around 17%) of the buried SOC can be expected to be preserved even 1000–1500 years after burial. On a short-term basis (5 years), farmyard manure buried to 60 cm depth has been observed to at least double the SOC content by preserving approximately 80% of the buried SOC in the subsoil (Egerzegi, 1959). This has been attributed to the lack of physical disturbance in deeper soil layers, e.g. by freeze-thawing, drying-rewetting cycles or regular tillage.

Although SOC content and stocks in forest and cropland subsoil were found to be enhanced over the long-term through deep ploughing, total SOC stocks in deep-ploughed forest soil were not greater than in reference soil. This is contrary to the observations made for deep-ploughed cropland soil. An important mechanism contributing to SOC sequestration in deep-ploughed soil is continuous SOC accumulation in the newly formed topsoil. The recovery and buildup of new SOC-rich topsoil was much slower in forest than in cropland soil.

The average time since deep ploughing was longer at the cropland sites studied here (on average 46 years, Table 3.1) than at the forest sites (on average 38 years, Table 3.1), resulting in a shorter SOC accumulation period for deep-ploughed forest topsoil. Nevertheless, even when accounting for the differences in time span, SOC stock differences in topsoil of deep-ploughed and reference soil were twice as high in forest soil as in cropland soil (Table A4.1). This may be due to the fact that in cropland soil, carbon inputs in the form of crop residues and leaf litter are directly incorporated into the mineral soil through regular tillage operations. In forest soil, on the contrary, aboveground litter first ends up in the forest floor and is thereafter partly transferred into the mineral soil (Fisher and Binkley, 2013). Because the forest sites studied here had acidic pH values (between 3 and 5) and *Pinus sylvestris* was one of the dominant tree species (Table 3.1), litter decomposition and SOC incorporation into the mineral soil has probably been slow and limited (Fisher and Binkley, 2013). This can be inferred from the fact that SOC stock differences in the L-horizon between deep-ploughed and reference soils were not significant while the F+H-horizon and topsoil had significantly less SOC in deep-ploughed than in reference soil (Fig. 3.2). This also indicated faster decomposition of litter in the deep-ploughed plots than in the reference plots, possibly related to less acid topsoil in the deep-ploughed plots because of admixture with previous subsoil material. However, the topsoil pH values were only slightly different in deep-ploughed and reference plots at the Schwenow site.

In the deep-ploughed forest topsoil, the amount of N needed to build up SOC stocks to a level comparable to that in reference topsoil would be between 0.09 and 0.74 Mg ha^{-1} (Supplementary Table 1). However, forests are usually not fertilised with mineral N, so the N input is mainly derived from atmospheric N deposition. For European forests, an average atmospheric N deposition of approximately 20 kg ha^{-1} yr^{-1} has been reported (Brumme et al., 2009). This shows that the maximum possible SOC sequestration in forest soil was restricted by the available N supply, whereas N restriction was less probable in cropland soil due to a 4- to 15-fold greater N input by mineral and organic fertilisers (see Table 2.2). The buildup of SOC stocks in the near-surface forest topsoil, which were lowered because of mixing with SOC- and N-poor subsoil material as a result of deep ploughing, thus encounters a certain N limitation compared with cropland topsoil. We hypothesised that under such relatively N-poor conditions, there was enhanced mineralisation by microorganisms of organic matter that reached forest topsoil in order to obtain N (N mining (Spohn, 2015)). Thus, N mining might have further slowed SOC accumulation.

Topsoil burial by deep ploughing increased SOC stability at all study sites (Fig. 3.3). Higher stability to mineralisation has also been observed on comparing SOC buried via depositional processes to reference surface soil (VandenBygaart et al., 2015). In the present study, we determined potential SOC stability via laboratory incubation under standardised temperature and soil moisture conditions. Thus, it can be concluded that the stability of buried SOC is not solely caused by environmental conditions at greater soil depth, such as temperature, oxygen or water limitations.

Selective preservation of certain SOC fractions with higher stability (Lehmann and Kleber, 2015) could theoretically explain the preservation of SOC that has been buried for several decades. The labile fractions of SOC would then be mineralised leaving the most stable fraction as buried SOC. However, this mechanism could not fully explain the results obtained in the present study, because great SOC loss after burial did not concomitantly result in increased stability of the remaining SOC (Fig. 3.6a). Contrary to our expectations, at three sites (Essemühle, Hemmelsberg and Schwenow), we observed only slight losses of buried SOC. At the same time, stability of buried topsoil SOC was 50-60% higher than that in reference topsoil. This high SOC stability is most likely related to the land use history as heathland or peatland (Table 3.1). Former heathland soils have been observed to contain very stable SOC possibly related to a high content of hydrophobic and toxic substances for decomposers (Alcántara et al., 2016; Jalal and Read, 1983; Sleutel et al., 2008). In contrast, at the Banteln, Drüber, Lindenburg and Rebberlah sites we observed an SOC loss of more than 50%, but only minor increases in SOC stability, with the maximum stability increase in the remaining SOC being 36%. These results underline that stability is not an intrinsic property of SOC, e.g. via poorly degradable compound classes, but might be controlled by environmental factors that are not yet fully understood for subsoil OC (C. Rumpel et al., 2012).

The fLF of SOC has previously been found to be the most easily mineralisable SOC fraction (Lützow et al., 2007), and it can be expected to be mineralised within one decade. However, our observations did not consistently show this trend, since at four out of 12 sites studied more than 50% of the fLF persisted in the buried SOC (Fig. 3.4). Surprisingly, we observed, on the one hand, increased stability of buried SOC compared with the reference topsoil OC without a concomitant loss of fLF (Fig. 3.6b). This is similar to previous observations in buried colluvial soil of Canada, which were found to contain equal or greater mass of LF carbon than surface soil (VandenBygaart et al., 2015). On the other hand, at eight out of 12 sites in the present study, fLF decreased upon burial by 50-80%, with only a slight increase in SOC stability of at most 42% (Fig. 3.6b). Thus, it can be concluded that the effect of burial on SOC stability is not driven by fLF loss. These findings support the conceptual model that SOC is continuously processed and mineralised by microorganisms and that no preferential mineralisation of a certain, more labile, SOC fraction occurs (Lehmann and Kleber, 2015).

Higher fLF content in buried topsoil stripes than in reference topsoil was observed at the Schwenow and Hemmelsberg sites (Fig. 3.6 and Supplementary Fig. 1). For Schwenow, a forest site, the high fLF could be attributed to abundant root biomass and related carbon inputs (Fig. 3.5b). At the same time, the Schwenow and Hemmelsberg sites showed a high degree of SOC preservation since burial (99% and 89%, respectively) and high SOC stability in the buried topsoil stripes

(37% and 51% higher than in the reference topsoil, respectively). Under the assumption that roots are the main source of fLF at both sites, these findings indicate that (i) additional carbon inputs from roots growing in buried topsoil stripes do not lead to additional loss of the buried SOC and (ii) the roots themselves may promote SOC storage in the subsoil. This is in line with findings that root-derived carbon persists over twice as long in soils before being decomposed than shoot-derived carbon (Rasse et al., 2005).

Roots grew preferentially in the buried topsoil stripes in the subsoil of deep-ploughed soil. We suggest that this is attributable to the higher organic matter content providing nutrients and water retention. This was confirmed by the visible presence of higher root biomass in deep-ploughed forest soil compared with reference soil (Fig. 3.5). Roots are a major source of subsoil SOC (Rasse et al., 2005; Tefs and Gleixner, 2012), both as exudates and in particulate form (Angst et al., 2016a). It has also been reported that subsoil loosening, another subsoil melioration option, promotes root proliferation into deeper soil layers (Cai et al., 2014). In the present study, this could only be confirmed for the forest sites because root biomass sampling at the cropland sites was conducted mainly during winter or after harvest.

The lower $\Delta^{14}C$ values observed for oLF and HF in buried topsoil stripes than in reference topsoil reflect the fact that input of fresh carbon into these soil fractions was drastically reduced due to burial. Cropland soils were dominated by carbon that was older than the nuclear weapon testing in the 1960s and 1970s (Trumbore et al., 2016) and recent carbon input from crop residues of the last years. In contrast, forest reference topsoil mostly displayed positive $\Delta^{14}C$ values (Fig. 3.7), indicating the dominance of nuclear test-derived carbon. This reflects the slow SOC exchange in the reference topsoil of forests compared with cropland so that $\Delta^{14}C$ values remain positive and are not "diluted" with newer atmospheric carbon assimilated by plants and transferred into the soil with a lower ^{14}C concentration and thus more negative $\Delta^{14}C$. The low ^{14}C content at the Viborg forest site underlines the former arable land use at this site just before deep ploughing.

In summary, deep ploughing can lead to increased SOC storage comprising two aspects: (i) greater stability of buried SOC and (ii) additional SOC accumulation in the "newly established" topsoil.

4 Legacy of medieval ridge and furrow cultivation on soil organic carbon distribution and stocks in forests

4.1 Abstract

Land management history can influence soil organic carbon (SOC) stocks over centuries. In this study, the impact of medieval ridge and furrow cropland management on SOC in forests was assessed. Continuous clockwise ploughing in rectangular fields moved topsoil from the outer part of strip-shaped fields towards the centre, thus forming a corrugated microtopography with peripheral furrows and central ridges. This tillage technique led to the burial of former topsoil under the ridges. The effect of this human-created microtopography and the centuries old topsoil burial on forest SOC spatial distribution and stocks was investigated.

Five sites with ridge and furrow field strips under deciduous forests on soils of differing texture in Germany were sampled, with three orthogonal transects of the field strips and a defined reference position where neither net soil removal nor accumulation occurred. Reforestation took place between the 17th and 19th century. At 0 to 10 cm depth, average SOC content was 28 $\pm$ 3 g kg^{-1} at ridges, 37 $\pm$ 3 g kg^{-1} at reference positions and 47 $\pm$ 5 g kg^{-1} at furrows. SOC stocks were 7 $\pm$ 5% lower at ridges and 8 $\pm$ 4% higher at furrows than at reference positions. Enhanced C input at furrows through leaf litter accumulation was indicated by higher SOC content in the free light fraction at furrows (10$\pm$5 g kg^{-1}) than at ridges (6$\pm$3 g kg^{-1}), higher specific SOC mineralisation (37 $\pm$ 4 µg CO_2-C g^{-1} SOC at furrows and 31 $\pm$ 3 µg CO_2-C g^{-1} SOC at ridges) and wider C/N ratio at furrows (18$\pm$1) compared with ridges (17$\pm$1).

Buried topsoil under ridges (20 to 33-52 cm depth) did not contain significantly less SOC than corresponding samples at furrows and reference positions. However, SOC content was 0.4 to 0.9 g kg^{-1} higher at ridges than at reference positions, indicating long-term preservation of former topsoil SOC by burial under ridges, although enhanced SOC stocks at ridges due to carbon burial could not be significantly confirmed for all sites. It can be concluded that preserved microtopography over centuries and ancient topsoil burial, a legacy of medieval ridge and furrow cultivation, still influences forest SOC spatial distribution and stocks.

4.2 Introduction

Human activity has been modifying ecosystems since the Neolithic Age through land management and land use changes (Price, 2000). These have affected the global carbon (C) cycle through either C losses to the atmosphere or C sequestration in biomass and soils. In particular, soil C dynamics do not respond rapidly to land cover change, but C stocks and fluxes may display a long-lasting effect (Houghton et al., 2012; Ciais et al., 2013). Primary emissions of 63 Gt C have been estimated to result from anthropogenic disturbance of the natural land cover in preindustrial times until 1850 arising mainly from Europe, India and China (Pongratz et al., 2009). By late medieval times, 4.6 x 106 km2 land world-wide, equivalent to 5% of the area potentially covered by vegetation, was under agricultural use (Pongratz et al., 2008) and atmospheric CO_2 exceeded its natural range of variation (Pongratz et al., 2009).

In Central and Western Europe, forest clearance for cultivation increased continuously between 1000 BC and industrialisation in the 19th century, with two main interruptions after the decline of the Roman Empire in the 5th century and the Black Death in the 14th century (Kaplan et al., 2009). During these times, many villages and their associated agricultural land were abandoned and forest was able to re-establish (Abel, 1955).

Ridge and furrow cultivation was a widespread, possibly dominant, agricultural practice in Europe during the Middle Ages and continued up to the 19th century (Eyre, 1955; Frank, 1912; F. S. Hartmann, 1882; Jäger, 1954; Mortensen, 1957; Trächsel, 1962). It consisted of ploughing a strip-shaped field clockwise using a simple mouldboard plough and draught animals. Through the repeated use of a ploughing pattern turning soil to the centre of the field (Epperlein, 1975), the topsoil towards the outer part of the field (furrows) was ploughed up and deposited toward the inner part of the field (ridges) (Fig. A5.1). After some decades, a corrugated land surface was formed, with a furrow in the outer part of the rectangular field and a convex ridge in the inner part. In this way, topsoil was buried under the ridges. Ridges were sometimes built up by hand prior to cultivation. Between the ridge and furrow positions, several decimetres height difference were built up (Meibeyer, 1969). The possible benefits of this land use practice were many, including (i) higher fertility of the ridges through organic matter accumulation by topsoil depth enlargement, (ii) a drainage effect of the furrows in wet soils, (iii) improved water storage in sandy soils, (iv) incorporation of unweathered subsoil containing nutrient-rich minerals and (v) establishment of clear field borders, thus minimising loss of topsoil to neighbouring fields (Meibeyer, 1977; Bartussek, 1982). However, since ridge and furrow fields were established on all types of soils, irrespective of drainage conditions, the most likely reason for their establishment was as a risk management strategy that helped to ensure a certain level of yield in dry (enhanced yields in furrows) and wet (enhanced yields on ridges) years. Crop growth was concentrated on the ridges (Bartussek, 1982). The fields were mostly used according to the three-field system rotating winter cereals (wheat or rye), summer crops (barley, oats or peas) and fallow for grazing (Küster, 1995).

Ridge and furrow fields were mostly located in the vicinity of villages (Niemeier, 1967). If a village was abandoned, cultivation of the fields was often abandoned too and the ridge and furrow microtopography was preserved over centuries under the subsequently established forest (Sittler, 2004). Some ridge and furrow fields have also been preserved under grassland (Beeresford and St Joseph, 1979). However, at most agricultural sites the ridges were levelled off in the late 19th and early 20th century with the growing use of earthenware tiles as subsurface drains and of drainage channels, as well as technological advances allowing deeper ploughing and thus higher accumulation of organic matter (Bartussek, 1982). The ridge and furrow legacy is thus not only present in abandoned fields under forest and grassland, but also in modern arable fields. Fossile, C-enriched Ap horizons have been observed at ridge and furrow sites under forest, reflecting the former agricultural history of these sites (Meibeyer, 1969; Niemeier, 1967; Well, 1989). However, consequences of this land use history on soil organic carbon (SOC) stocks are completely unknown. The microtopography may have influenced the spatial distribution of SOC over time, for example via preferential accumulation of litter in the furrows or effective long-term SOC storage under the ridges.

With the development of ridge and furrow systems and the anthropogenic creation of microtopography, SOC-rich topsoil material was buried deeper than the recent forest Ah horizon. Carbon burial has been suggested as an effective measure to store additional SOC in subsoils on a long-term basis (N. T. Chaopricha et al., 2014; Hoffmann et al., 2013; Kristof Van Oost et al., 2012; VandenBygaart et al., 2015). Translocation of SOC to greater soil depth has been described as a promising SOC sequestration measure due to the low SOC contents in subsoils with high apparent SOC ages, their high content of unsaturated mineral surfaces and their consequently assumed high SOC storage capacity (Baldock and Skjemstad, 2000; Beare et al., 2014). Large-scale SOC storage through burial has been reported in depositional footslopes (VandenBygaart et al., 2015), volcanic deposits (Basile-Doelsch et al., 2005) and deep-ploughed arable soils (Alcántara et al., 2016). In this study we investigated the legacy of medieval ridge and furrow cultivation on the forests that now occupy these sites, including the effects of topsoil burial and the impact of the microtopography on SOC. The following questions were addressed:

1. Does ridge and furrow microtopography lead to preferential litter input into furrows, and thus shape the distribution of recent SOC stocks under forest?
2. Does topsoil burial under ridges result in higher SOC stocks compared with reference positions caused by high stability of the buried SOC?

4.3 Materials and Methods

Selection of study sites and soil sampling

Five forest sites with a documented and approximately dated ridge and furrow land use history were selected for sampling. All sampled ridge and furrow fields were associated with former villages that were deserted between the 13th and 16th century. We searched in particular for the most suitable sites with respect to historical documentation, as much time as possible since abandonment, a high degree of conservation of the ridge and furrow systems and a relief height difference of at least 30 cm between ridge and furrow. A soil texture range from loamy sand to silty loam was covered (Table 4.1). The five sites were located in northern and central Germany.

Sampling design was based on the ridge and furrow description according to Meibeyer (1969). The highest position within the undulating rectangular field strip was identified as the ridge and the lowest as the furrow (Fig. 4.1). At a certain position within the transverse profile of the rectangular field strips, the elevation of the soil surface remained unchanged and neither net topsoil excavation nor net accumulation took place (Fig. A5.1). This was taken as the reference position and was assumed to be located at 57% of the distance between ridge and furrow, based on the shape of ridge and furrow systems documented by Meibeyer (1969). He determined a reference position on the original land surface as the position at which the soil mass (measured as drawn area of a ridge and furrow profile) lacking at the furrow was equal to that accumulated at the ridge. Sampling at each site comprised two ridges, two furrows and four reference positions. In addition, a position between ridge and reference (middle 1, Fig. 4.1) and a position between reference and furrow (middle 2, Fig. 4.1) were sampled. Samples were collected to 100 cm depth at the ridge positions and the equivalent depth at the middle 1, reference, middle 2 and furrow positions.

Ploughing depth was assumed to be a maximum of 20 cm (Meibeyer, 1969). The deepest limit of the buried topsoil under the ridges was marked in the field by fixing of a string at two reference positions to each side of the ridge, at 20 cm depth, in order to obtain a straight line parallel to the original surface and marking the maximum depth of initial ploughing. At the ridge position, the depth increment between 20 cm and the string was taken as the buried topsoil (increment 3), which was sometimes greater than 20 cm because its thickness depended not only on the maximum ploughing depth but also on the amount of accumulated soil at the ridge. The following depth increments were thus sampled (Fig. 4.1): (1) recent Ah horizon developed under forest within the former plough layer (0 to 10 cm), (2) last active ploughed layer down to maximum ploughing depth (10 to 20 cm), (3) buried topsoil as relictic Ap horizon (ridge and middle 1: 20 to 33-52 cm depth), (4) non-tilled subsoil with equivalent depth of buried topsoil horizon at ridge (ridge and middle 1: 33-52 to 46-84 cm depth, reference, middle 2 and furrow: 20 to 33-52 cm depth) and (5) down to 100 cm at ridges. At the Eddessen site, the last 20 cm (80 to 100 cm) were sampled separately because a pedogenic horizon with stagnic properties and high clay content was identified.

TABLE 4.1: Characteristics of the ridge and furrow study sites

Site	Latitude, Longitude	Elevation (MAMSL[1])	MAT (°)	MAP (mm)	pH[2]	Sand (%)	Silt (%)	Clay[3] (%)	Soil Type[4]	Parent material
Bad Helmstedt	52°13'94"N 11°3'12"E	165	9.4	608	4.7	64	27	9	Stagnic Cambisol	Pleistocene sandy till over loamy till
Eddessen	51°36'15"N 9°18'54"E	280	8.9	799	3.2	4	83	13	Haplic Stagnosol	Triassic claystone covered by Pleistocene loess
Hohnstedt	52°22'45"N 10°42'51"E	110	9.7	685	3.3	76	17	7	Stagnic Cambisol	Pleistocene sandy till over black Jurassic claystone
Kreuzheide	52°27'32"N 10°46'54"E	75	9.7	635	4.0	77	15	8	Dystric Cambisol	Pleistocene sandy till over loamy till
Reinhausen	51°26'22"N 10°0'19"E	345	8.1	747	4.8	8	76	16	Stagnic Luvisol	Weichselian loess overlying Triassic sandstone

[1]Metres above mean sea level
[2]As determined from the reference Ah-horizon
[3]Texture values from the reference Ah-horizon
[4]According to IUSS Working Group WRB (2015)

Soil cores with a fixed volume were retrieved from three parallel transects 15 m apart, which were set orthogonal to the ridge direction. A soil coring probe with 60 mm inner diameter, driven by an electric jackhammer (Wacker EH 23, Wacker Neuson, Munich Germany), was used for sampling, as recommended by Walter et al. (2016). Each transect comprised two ridges, four middle 1 positions, four reference positions, four middle 2 positions and two furrows. Thus, a total of 16 soil cores per transect and 48 soil cores per site were sampled. All three cores from each position formed one composite sample per depth increment. All vertical and horizontal distances between positions of the cross-section ridge and furrow field along the sampling transects were documented.

In addition to the transect sampling, a 20 m long and 1 m wide cross-sectional transect was excavated using a mini-digger. Disturbed samples were extracted from the soil profile for chemical analyses and incubation experiments. These samples were stored frozen at -19°C until later analysis. Sampling was conducted between May and July 2014.

Sample preparation and laboratory analyses

Core samples and samples from the soil profile were dried at 65°C to constant mass and sieved to <2 mm, the fraction used for analysis. Loamy samples were coarsely crushed to 2 mm in a jaw crusher (BB1; Retsch, Haan, Germany) instead of sieving. Stones and coarse roots removed with sieving were weighed and their mass was subtracted from the total soil mass to obtain the fine soil mass. Fine roots were removed manually during sample preparation in the laboratory. Aboveground residues and the forest floor were removed directly in the field prior to sampling of soil cores.

Soil texture was determined by the sedimentation method (Moschrefi, 1983). The pH was measured in 0.01 M $CaCl_2$ using a glass electrode. The cations Ca^{2+}, Mg^{2+}, K^+, Na^+ and Al^{3+} were extracted from 100 g soil with 0.1 M $BaCl_2$ (Kretzschmar, 1996) and quantified using atomic absorption spectrometry (AAS) (Perkin Elmer AAS 4100, Rodgau, Germany) to determine cation exchange capacity (CEC). Dithionite-soluble Fe (Fe_D), as a measure of crystalline iron, was extracted by duplicate reduction with $Na_2S_2O_4$ at 85 °C (Mehra and M. Jackson, 1958). Amorphous Fe and aluminium (Al) oxides (Fe_O, Al_O) were obtained by extraction with a mixture of ammonium oxalate and oxalic acid in darkness. Concentrations of Fe and Al in the dithionite and oxalate extracts were determined by AAS (AA-280FS, Varian, Palo Alto, CA, USA).

An aliquot of each sieved sample was milled in a planetary ball mill and analysed by dry combustion for total C and total nitrogen (N) (TruMac CN LECO, St. Joseph, MI, USA). All soils were entirely decalcified (pH<6), so no determination of inorganic C was conducted. The SOC content was expressed as g SOC kg^{-1} soil dry matter. For the evaluation of SOC biodegradability, incubation experiments were conducted over 24 hours with three replicate field-fresh samples at 40-50% of their water-holding capacity (Martens, 1995). The incubation was performed at 22°C, to determine short-term SOC mineralisation rates, using a semi-automatic continuous flow system (Heinemeyer et al., 1989). Since the samples had been stored in the freezer, a pre-incubation period of one week at 22°C preceded analysis. Evolving CO_2 was measured hourly with an infra-red analyser (ADC-255-MK3, Analytical Development, Hoddesdon, England). SOC mineralisation was calculated as the mean of relatively constant CO_2 production rates in the second half of the incubation and expressed as specific SOC mineralisation rate (µg CO_2-C g^{-1} SOC h^{-1}).

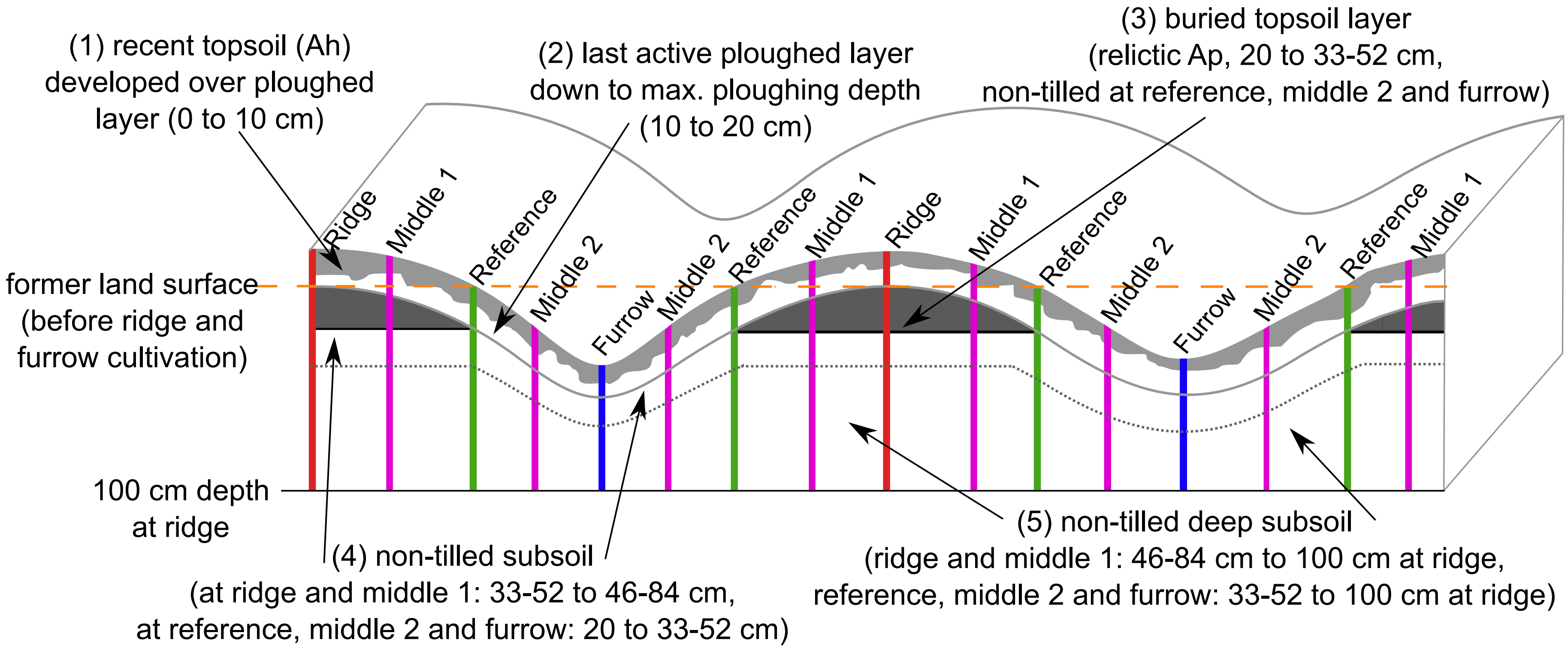

FIGURE 4.1: Schematic cross-sectional ridge and furrow profile and sampling design. The dashed yellow line shows the original land surface before ridge and furrow cultivation. Hereunder lays the buried topsoil. In the course of ridge and furrow ploughing, a furrow was developed. At the same time, soil accumulated in the central part of the land strips forming a ridge. In the ridge and furrow fields, there is a position where there was no net soil removal or accumulation, this position served as a reference.

Carbon input and SOC stabilisation through adsorption to minerals was estimated by SOC density fractionation of 30 g sieved soil dry matter (Cerli et al., 2012; Golchin et al., 1994). Free light fraction (fLF) was separated after addition of 120 ml 1.6 g cm^{-3} sodium polytungstate solution (SPT 0, TC Tungsten Compounds, Grub am Forst, Germany). Occluded light fraction (oLF) was obtained after resuspension in SPT and dispersion by ultrasound with an energy input of 400 J ml^{-1} calibrated according to Schmidt et al. (1999). Filtered fLF and oLF were rinsed with deionised water until the conductivity of the wash water was <10 µS cm^{-1}. Heavy fraction (HF) remained as sediment and was washed three times with 150 ml deionised water. All fractions were dried at 40°C before weighing and determination of SOC content. The SOC content of fLF is expressed as g SOC in the fraction per kg soil dry matter, while that in HF is expressed as relative amount of total SOC contained in the fraction.

Calculations and statistics

Differences in SOC content between the ridge, reference and furrow positions were tested with a linear mixed effects model fitted by restricted maximum likelihood using package nlme in R (Pinheiro et al., 2015), with the cross-section position as a fixed effect with three levels (ridge, reference and furrow) and site as a random effect. Three separate tests were performed for the first three soil depth increments (0 to 10 cm, 10 to 20 cm and the buried topsoil layer). The SOC contents determined in core samples and profile samples were included. The variances were weighed by increasing SOC content with the *varConstPower* function. Homoscedasticity of the standardised residuals was confirmed visually. The P-values (P) were extracted from the model output showing results from t-statistics. In the text, the degrees of freedom (df) and the t-value (t) are also given.

Mass-equivalent SOC stocks were calculated with the lightest soil core at a furrow position of each site as the reference mass (B H Ellert et al., 2007). SOC stocks were calculated as Poeplau and Don (2013):

$$\text{SOC stock} = \sum_{i=1}^{n}\left(\frac{\text{FSM}_i}{\text{V}_i} \times \text{C}_i \times \text{D}_i\right)$$

where FSM is the fine soil mass (sampled soil mass minus root and stone mass) at each depth increment i, C_i is the SOC content and D_i is the height of the depth increment over n depth increments down to the depth of the lightest furrow. SOC stocks were calculated as Mg SOC ha^{-1} and then normalised so that the average stock at the reference positions (N=4) represented 0%, in order to enable comparisons between all sites. The significance of the SOC stock differences between the cross-section positions was tested with a linear mixed effects model fitted by restricted maximum likelihood with position as a fixed effect with five levels (ridge, middle 1, reference, middle 2 and furrow) and site as random effect. In this case, variances were weighed by increasing relative SOC stock at each site with the *varPower* function to ensure homoscedasticity.

A comparison between SOC stocks in the ridge and furrow systems and a hypothetical system represented by the reference positions was conducted to evaluate the ridge and furrow legacy. Each ridge and furrow field was separated into two halves, each comprising one ridge with two middle 1 positions, two references, two middle 2 positions and one furrow. In this way, the variability within each site could be quantified as a standard error. The mass equivalent SOC stock was multiplied by a weighing factor according to the area covered by each sampling position. All values given in the text are arithmetic means with standard error, unless otherwise stated. Differences, tested with paired Wilcoxon rank sum tests, and correlations, tested using Spearman's rho, were significant (=0.05) only when explicitly stated. All statistical analyses were performed with R version 3.2.3 (R Core Team, 2015).

Land use history of the study sites

The five study sites covered a wide range of soil types (Table 4.1). All sampled soil profiles are shown in Fig. 4.2. The land use history of the sampled ridge and furrow sites was also very diverse (Table 4.2). The Bad Helmstedt site is situated in a forest site defined by the forestry district in charge as "historical old forest", i.e. at least 100 years old. The ridge and furrow area is associated with the Slavic village Bemesdorf, which was deserted at least by 1224. In that year, Count Palatine Heinrich gave the area to the Augustinian nunnery of Marienberg as a gift, which included forest, grasslands and cropland (Jarck, 1998). The convent of Berge in Magdeburg also owned part of the old Bemesdorf area as agricultural land (Meier, 1896). In the historical land survey of the Land Brunswick from 1764-1784, the area is marked as forest still belonging to the Marienberg nunnery. The ridge and furrow system comprises an area of 16 ha and is among the best preserved in the region.

The ancient village related to the ridge and furrow complex at the Eddessen site was first mentioned in the year 1003. Abandonment of the village began in the 14th century and continued until the 17th century. In 1368 Eddessen was sold to the noble family von Spiegel, to whom the land still belongs today. The main devastation of the village was due to fighting in 1444-1447 in the Soester feud. In 1656, the area was documented as "silva Eichhagen" (Bergmann, 1991). In the 17th century an anchoritage was established at the same location, which was inhabited until the 20th century. There are 82 ridges conserved under the Eichhagen forest, covering an area of almost 30 ha (Bergmann, 1991).

Over 60% of the Hohnstedt forest has a ridge and furrow history, covering 174 ha (Krutsch, 1966). The first documentation of the existence of the former village Hohnstedt dates back to 1322, with the construction of a church (Krutsch, 1966) and a record of the regional bishop demanding tithe payments from the farmers (Behrends, 1849). By 1553 the village had been abandoned, but it is unclear whether the field sites were still cultivated and the families resettled in nearby villages. In the land survey of the Electorate of Hanover from 1781, the entire area is marked as deciduous forest.

The ridge and furrow fields at the Kreuzheide site belonged to the deserted village Badekot, which was first mentioned around 1160 in the estate list of the convent "S. Lugdgeri von Helmstedt" (Meier, 1896; Mutke, 1913). In a field map from around 1300, the sampled area is still marked as a ridge and furrow field as part of the already abandoned village. Another mention of the deserted village of Badekoth is made in documents from 1476 stating that abandonment was probably caused by war (Kleinau, 1967). The area is marked as part of the deserted village Badekot in the historical land survey of the Land Brunswick from 1764-1784. At this time the ridge and furrow field was either still being cultivated by farmers in nearby villages (Meibeyer, 1977) or had developed into heathland, which extended north and south of the deserted land boundary. In 1899 the area was mapped as coniferous forest in the Prussian land survey. The ridge and furrow complex comprises approximately 30 ha.

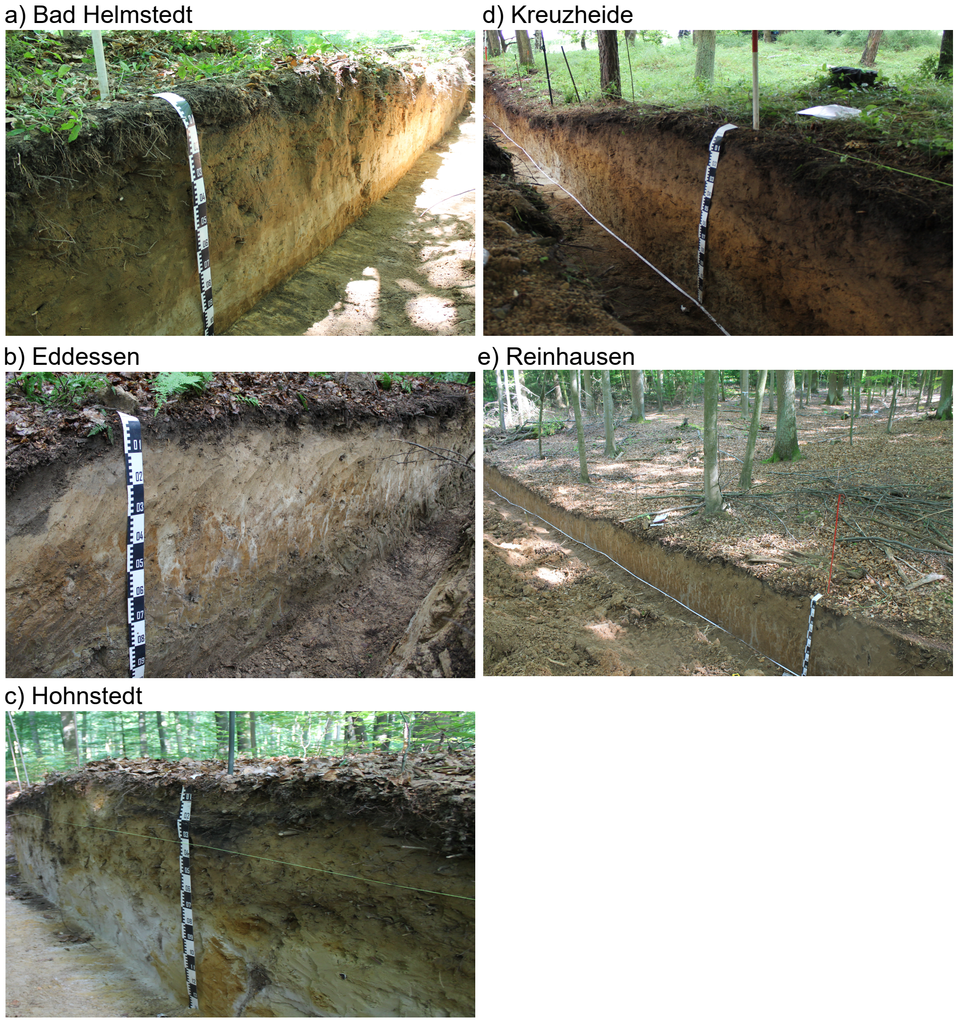

FIGURE 4.2: Sampled cross-sectional soil profiles. Ridge positions are marked by the measuring tape. Samples were collected down to 100 cm at the ridge position. Furrow position was determined as the lowest position.

The total ridge and furrow system at the Reinhausen site covers around 8 ha and consists of 28 cultivated ridges (Well, 1989). The ridge and furrow system was most likely developed in the 12th and 13th century, when large areas of the Reinhausen forest were deforested and the use of mouldboard ploughs became common in Europe. A settlement Scrozinroth belonging to the ridge and furrow system was first mentioned in 1118 (Kühlhorn, 1994) and later in 1207 in the archives of the Reinhausen convent as part of the estate (Hamann, 1991), where it was still inhabited. Historical estate lists and sales agreements from 1459 and 1464 indicate that the land belonging to Scrozinroth had stopped being cultivated and a forest was established (Well, 1989). The settlement does not appear in the estate inventory of 1508. In a map of 1595 from the public records office of Hanover, the area is marked as forest. A new wave of deforestation and cultivation of the area probably took place at the beginning of the 18th century, because in maps from that time the area appears as an agricultural field. In the land survey of the Electorate of Hanover from 1785, the area is marked as a cultivated field within the surrounding forest. Reforestation was undertaken in 1872, which agrees with the age of the *Quercus robur* that constitutes the main tree species in this forest (Table 4.2).

TABLE 4.2: Management and land use history of the ridge and furrow study sites

Site	Tree species, age and proportion of area	Starting year of ridge and furrow cultivation[5]	Year of reforestation	Mean height between ridge and furrow (m)	Mean horizontal distance between ridge and furrow (m)	Mean slope between ridge and furrow (%)
Bad Helmstedt	*Fagus sylvatica*, 65 yr, 95% *Larix decidua*, 57 yr, 5%	earlier than 1224	latest 1784	0.52	7.4	8%
Eddessen	*Fagus sylvatica*	1003	1656	0.31	8.8	4%
Hohnstedt	*Fagus sylvatica*, 143 yr, 50% *Quercus robur*, 143, 35%; 114 yr, 15%	1322	latest 1781	0.37	5.0	9%
Kreuzheide	*Pinus sylvestris*, 151 yr, 100% *Quercus petraea*, 63 yr, 85%[6]	1160	latest 1899	0.56	5.3	10%
Reinhausen	*Quercus robur*, 142 yr, 90% *Fagus sylvatica*, 73 yr, 10%	1118	1872	0.50	7.6	7%

[5] Oriented on first document mentioning associated village
[6] In the understorey

4.4 Results

Depth distribution of SOC content in the ridge and furrow profile

Mean topsoil SOC content significantly increased in the cross-section profile in the order ridge, reference and furrow (Fig. 4.3). The largest gradients were observed in the topsoil (0 to 10 cm depth), with ridges containing an average SOC content of 28 ± 3 g kg^{-1}, reference positions 37 ± 3 g kg^{-1} and furrows 47 ± 5 g kg^{-1} (Fig. 4.3). Differences were significant between ridge and reference (df=48, t=2.5, P=0.01) and between ridge and furrow (df=48, t=3.6, P<0.01), but not between reference and furrow (df=48, t=1.6, P=0.1).

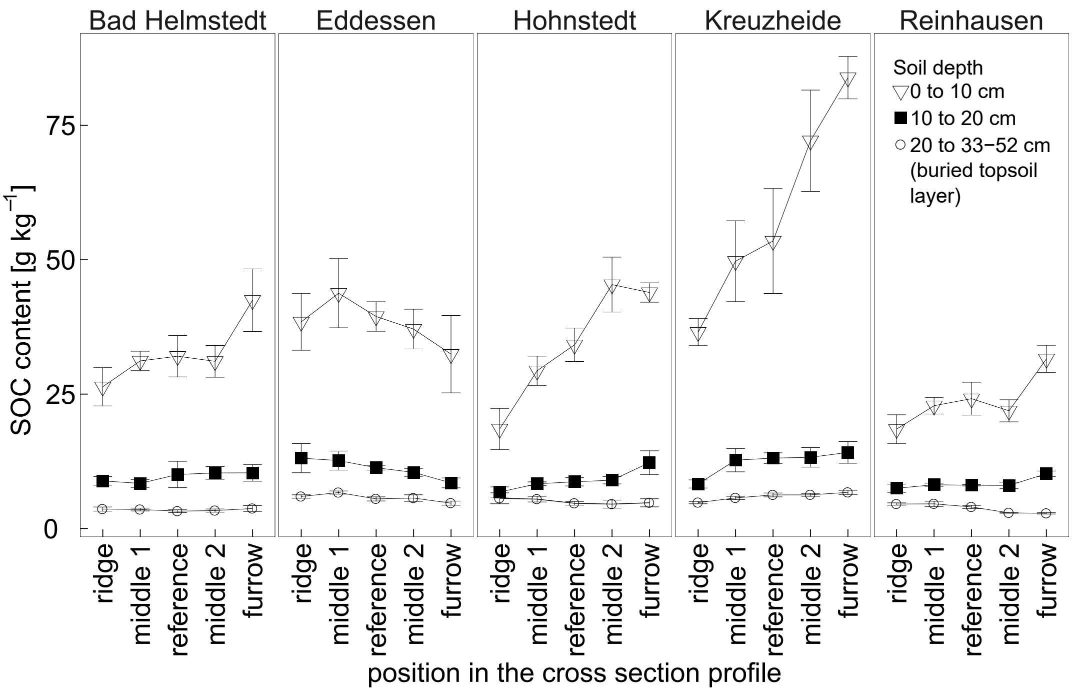

FIGURE 4.3: SOC content at the different positions of the cross-sectional ridge and furrow profiles. Middle 1 position was between ridge and reference, middle 2 position was between the reference and furrow. Shapes represent arithmetical means and bars their standard errors (N= 3 at ridge and furrow, N=5 at reference, N=4 at middle 1 and 2).

The cross-section gradient in the topsoil was most pronounced at the Kreuzheide site, where the SOC content at ridges was 37 ± 3 g kg^{-1}, at reference positions 53 ± 10 g kg^{-1} and at furrows 84 ± 4 g kg^{-1} (Fig. 4.3). The microtopography was also most pronounced at the Kreuzheide site, with a mean height difference between ridge and furrow of 56 cm (Table 4.2). Eddessen was the only site with an opposing trend, with SOC content at ridges being 38 ± 5 g kg^{-1} and at furrows 32 ± 7 g kg^{-1}. The absolute SOC content at furrows at 0 to 10 cm depth was significantly correlated with the slope between the two positions (Rho=0.7, P<0.01, Fig. 4.5a). Moreover, the SOC content at ridges was negatively correlated with the slope from ridge to furrow (Rho=-0.06, P=0.8, Fig. 4.5a).

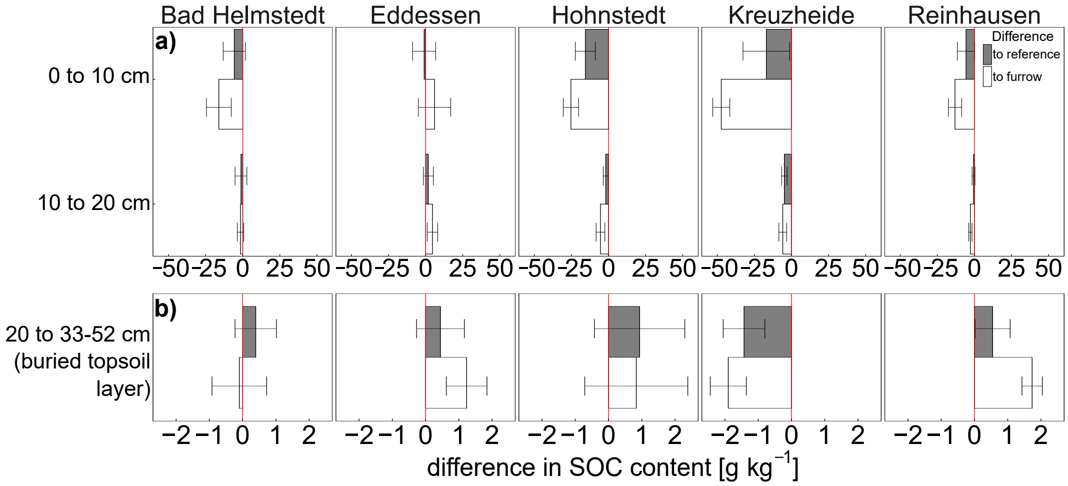

FIGURE 4.4: Differences in SOC content (ridge minus reference position in grey, ridge minus furrow in white) for a) first two depth increments and b) buried topsoil layer and same depth at furrow and reference. Bars represent stand errors of the mean.

The trend in the depth increment of the buried topsoil was the opposite to the observed trend in the topsoil (Fig. 4.3). The average SOC content of the buried topsoil at ridges (4.9 ± 0.3 g kg^{-1}) and middle 1 positions (5.2 ± 0.3 g kg^{-1}), was almost equal to or slightly higher than at reference positions (4.8 ± 0.3 g kg^{-1}), at the same soil depth even though differences were not significant. The difference in SOC content in the buried topsoil layer between ridges and reference positions ranged from 0.4 ± 0.6 g kg^{-1} at Bad Helmstedt to 0.9 ± 1.4 g kg^{-1} at Hohnstedt (Fig. 4.4b).

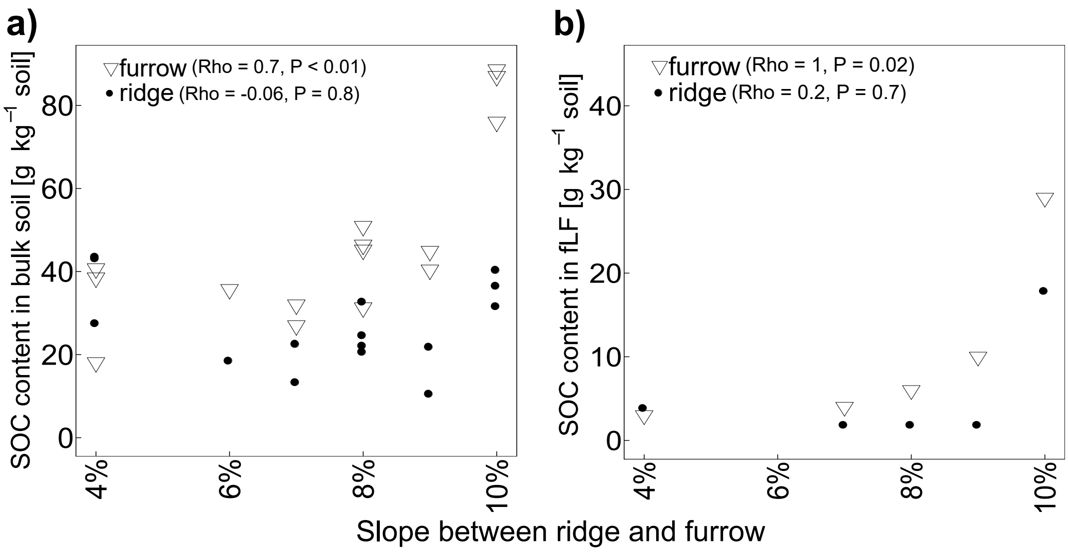

FIGURE 4.5: Relationship between the ridge slope and the SOC content at 0 to 10 cm depth in **a)** bulk soil and **b)** fLF.

Other soil properties along the ridge and furrow transects

Ridge and furrow cultivation could have influenced not only the depth distribution SOC, but also other soil properties that may determine SOC turnover and stabilisation. Therefore the relationship between differences in SOC concentrations and differences in other soil properties were assessed (Fig. A5.2). The strongest correlations to the differences in topsoil SOC content between ridge and furrow were found for the differences in N content (Rho=0.9, P=0.08), cation exchange capacity (Rho=0.7, P=0.2) and clay content (Rho=0.8, P=0.09) but none of the relations were significant.

Similarly, the differences in the soil properties of the buried topsoil compared with those of the same depth at reference positions were not consistently related to SOC differences. The only correlations observed were for N content (Rho = 1, $P < 0.02$) and amorphous oxides (Rho = 0.9, P = 0.08) (Tab. A.1).

SOC stocks in the cross-section of the ridge and furrow profiles

On average, the SOC stocks at ridges were 6±5% lower and at middle 1 positions 3±3% lower than at reference positions (Fig. 4.6; Tab. A.2). In contrast, the SOC stocks at middle 2 positions were on average 1±3% higher than at reference positions and those at furrows were 10±4% higher. The differences were significant between ridges and furrows (df = 33, t = 3.7, $P < 0.01$) and between reference positions and furrows (df = 33, t = 5.8, $P < 0.01$), but not between ridges and reference positions (df = 33, t = -0.8, P = 0.4). On average, 53 to 73% of total SOC down to 54 cm (Table 4.2) was stored from 0 to 20 cm depth.

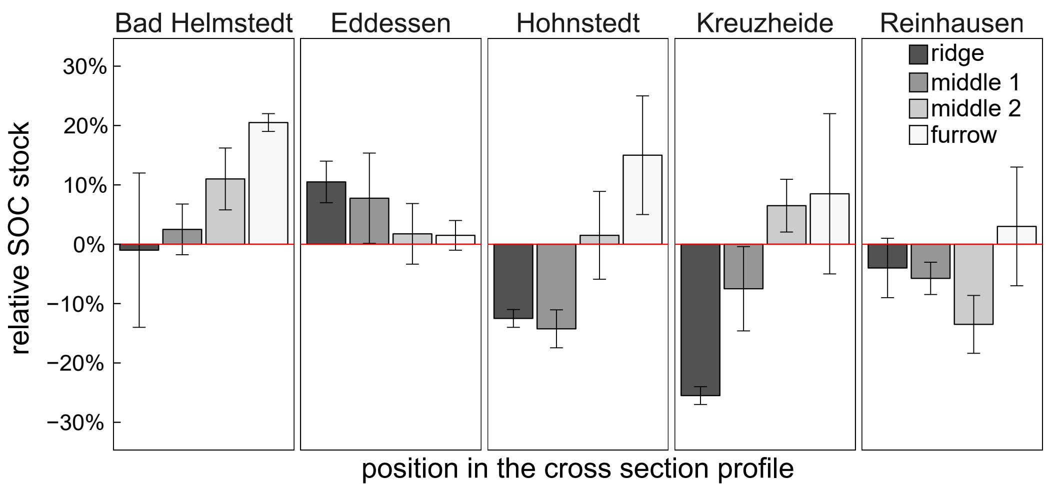

FIGURE 4.6: SOC stocks relative to the reference SOC stock. Arithmetical means of the transect core samples are shown. Bars represent standard errors of the mean (N=2 at ridge and furrow, N=4 at middle 1 and 2 positions).

The largest SOC stock difference between ridge and furrow with the reference as a baseline (35±7%) was observed at Kreuzheide. Eddessen was the only site with higher SOC stocks at ridges compared with the reference positions (10±3%) and lower SOC stocks at furrows (1±5%). The SOC stocks at ridges relative to the reference position were negatively correlated to the slope between ridge and furrow (Fig. A5.3).

Total SOC stocks in the ridge and furrow system were not significantly different from SOC stocks at a hypothetical reference position system (Fig. 4.7). At the Bad Helmstedt, Eddessen and Hohnstedt sites, the ridge and furrow system had on average 8%, 5% and 2% higher SOC stocks, respectively, than the reference only scenario. However, at the Kreuzheide and Reinhausen sites, the SOC stocks in the ridge and furrow systems were 4% lower than those at the hypothetical reference position system.

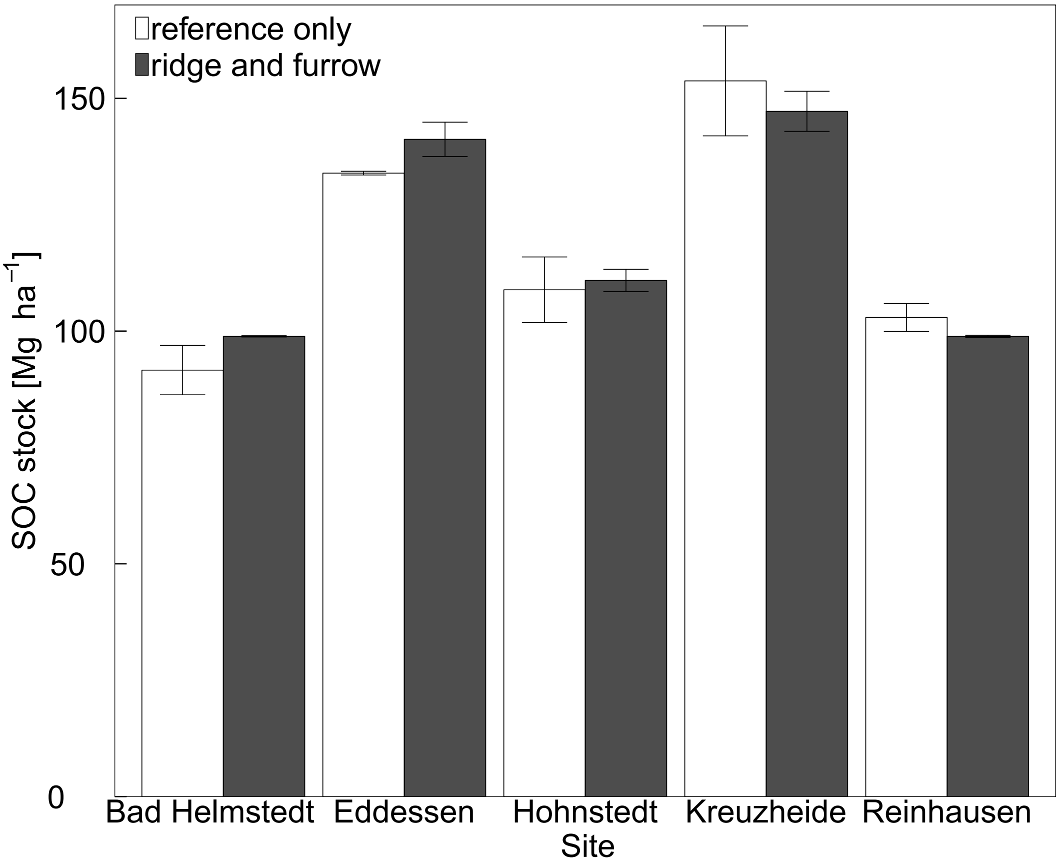

FIGURE 4.7: Field scaled SOC stocks of the ridge and furrow sites compared to a hypothetical reference without ridge and furrow microtopography.

SOC quality along the ridge and furrow cross-section

In the topsoil (0 to 10 cm depth), more SOC in fLF was found at furrows (10±5 g kg^{-1}) than at reference positions (8±3 g kg^{-1}, P = 0.1) or at ridges (6±3 g kg^{-1}, P = 0.5). However, the error was large, indicating great variability between the sites and resulting in differences not being significant. The highest SOC content in fLF was observed at furrows at the Kreuzheide site (29 g kg^{-1}). A high and significant correlation was found between the SOC content in fLF at furrows and the slope between ridge and furrow (Fig. 4.5b). The proportion of SOC in HF increased with soil depth (Fig. 4.8). At the reference position in 0 to 10 cm depth, 47±7% of total SOC was found in HF, while at 22 to 33-52 cm depth the proportion was 70±4%. Similarly at ridges, the proportion of HF in total SOC was 53±11% at 0 to 10 cm depth and 73±5% in the buried topsoil layer (22 to 33-52 cm).

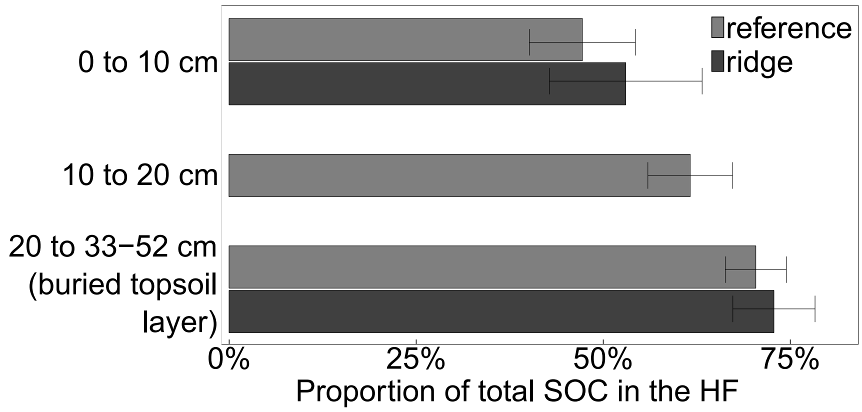

FIGURE 4.8: Relative proportion of total SOC in HF in different depths at reference and ridge. Bars with standard errors represent mean values throughout all sites (N=5).

Short-term specific topsoil SOC mineralisation (Fig. 4.9) was on average lower at ridges (31 ± 3 µg CO_2-C g^{-1} SOC) than at reference positions (37 ± 5 µg CO_2-C g^{-1} SOC) or furrows (37 ± 4 µg CO_2-C g^{-1} SOC), but the differences were not significant. Buried topsoil at ridges (16 ± 2 µg CO_2-C g^{-1} SOC) had lower specific SOC mineralisation compared with the same depth at reference positions (18 ± 2 µg CO_2-C g^{-1} SOC, P=0.1) or furrows (21 ± 3 µg CO_2-C g^{-1} SOC, P=0.02).

At 0 to 10 cm depth, C/N ratio (Fig. A5.5) was on average narrower at ridges (17±1) than at reference positions (18±1) and significantly narrower than at furrows (18±1). The C/N ratio in buried topsoil at ridges (13±1) showed no significant differences compared with the corresponding depths at reference positions (14±1, P=0.8) and furrows (14±1, P=0.2). Non-tilled subsoil at ridges (33-52 to 46-84 cm) had the lowest C/N ratio (12±1), but not significantly lower than in the buried topsoil layer (P=0.1).

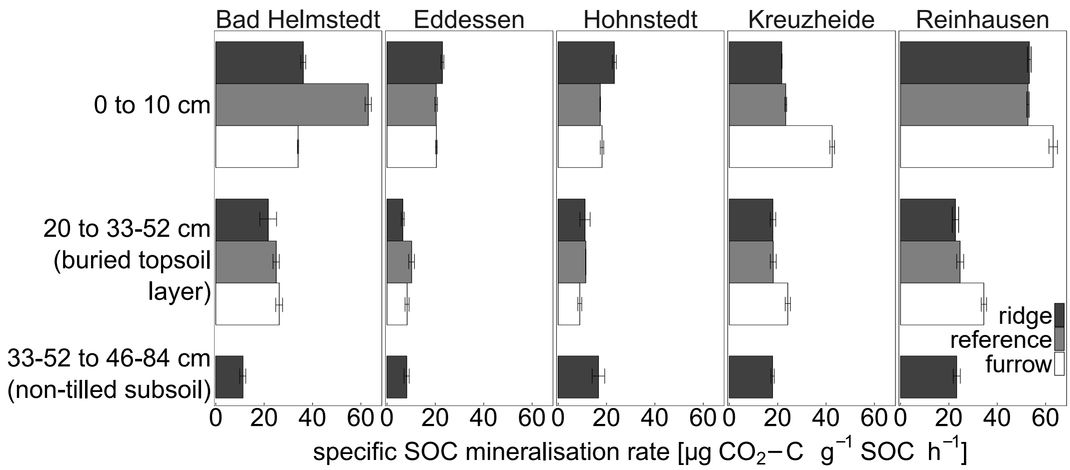

FIGURE 4.9: Specific SOC mineralisation rate of samples from different soil depths at ridge, reference position and furrow. Bars represent standard errors of the mean.

4.5 Discussion

Topsoil SOC content and total SOC stocks are influenced by ridge and furrow microtopography

The SOC content at furrows was closely related to the slope between ridge and furrow (Fig. 4.5a). Above a slope of 7%, SOC content was enhanced compared with at ridges. This suggests lateral C transport from ridge to furrow through enhanced C input with leaf litter in furrows. The Kreuzheide site had the steepest slope (Table 4.2) and also the greatest difference in SOC content between ridge, reference and furrow (Fig. 4.4). Moreover, the ridge and furrow system at Kreuzheide was located close to the forest edge, facing prevailing wind direction from the west. It can thus be expected that litter was redistributed by westerly winds from ridge into furrow. The SOC content in fLF at furrows, an indicator of litter C input, was strongly enhanced in ridge and furrow systems with a steep slope (Fig. 4.5b). With increasing slope, SOC content in fLF at ridges was lower than at furrows. In addition, C/N ratio at 0 to 10 cm at ridges was significantly narrower than at furrows (Fig. A5.5), which indicates that SOC at furrows was younger and less transformed than at ridges. The C/N ratio of older, highly transformed SOC is narrower because C is respired while N is relatively enriched in the course of decomposition and transformation (Conen et al., 2008). Enhanced C input into furrows compared with ridges is less likely to have occurred through dissolved organic carbon interflow because at the study sites with higher SOC content at furrows there was no impermeable subsoil layer.

Erosion related to the ridge and furrow microtopography is not likely to have influenced SOC content and stocks. Erosional displacement of soil from ridges to furrows has been observed in preserved ridge and furrow systems in Cumbria, England, but only under grassland (Cannon and Reid, 1993). Indicators of this were higher porosity, higher levels of exchangeable potassium and higher pH at furrow than at ridge, which also suggests that annual fertiliser was washed down from ridge to furrow. Erosion towards furrows (channels) from raised fields with a steeper slope has been observed in the Bolivian Lowlands under savannah vegetation (Rodrigues et al., 2015). However, due to the height difference between ridge and furrow of 120 to 160 cm in that study, erosion rates were much higher than in our study sites under forest, where the ridge to furrow height difference was 31 to 56 cm (Table 4.2).

One of the reasons for practising ridge and furrow cultivation was the drainage effect of ridges (Kerridge, 1951). The microtopography promotes preferential surface runoff towards furrows causing higher soil pore water contents at furrows and also litter transport towards furrows. Studies on the distribution of grass species at the Kreuzheide site have indicated higher moisture at furrows. Thus Deschampsia flexuosa, a grass that preferentially grows on dryer soils, dominates on ridges, while Molinia caerulea, which preferentially grows on more moist soils, dominates on furrows (Meibeyer, 1969). Similarly, in Cumbria, England, grass species have been observed to be significantly related to the ridge and furrow microtopography in an ancient meadow (Cannon and Reid, 1993). Species that prefer moist soil conditions preferentially grew on the furrows, which was related to higher volumetric water contents measured at furrows than at ridges. In Britain, ridge and furrow systems were created in forests to promote drainage and avoid waterlogging, and thus improve tree growth conditions (Moffat and Roberts, 1989).

In the west coast of New Zealand, which is a high rainfall area, soils with iron pans are dug over to create surface drains forming 3 m high and 45 to 50 m wide “humps and hollows” (Horrocks et al., 2010), which can be regarded as a modern version of the ridge and furrow system. The main aim of that land modification is to enhance surface runoff and reduce waterlogging. In the short-term after land modification, SOC accumulates faster in the humps than in the hollows, which is most likely related to higher productivity on humps and enhanced runoff and nutrient leaching in hollows (Thomas et al., 2007). Similar effects probably arose in ridge and furrow fields during cultivation in wet years. After reforestation, plant growth was probably no longer greater at ridges, so that the effect of higher litter input at furrows could determine SOC dynamics over the centuries. Immediately after field abandonment, SOC probably accumulated at a fast rate, especially at furrows. In abandoned fields with subsequent secondary forest expansion, where no subsoil was exposed to surface C input, pre-agricultural levels of C in the topsoil are predicted to have recovered within 200 (Kalinina et al., 2013) or 230 years after abandonment (Knops and Tilman, 2000). For our study sites (approximately between 115 and 358 years old, Table 4.2), topsoil SOC content can be expected to be no longer diminished due to historical agricultural land use.

At the Eddessen site, we observed an opposing trend, higher SOC content at ridges than at furrows over 0 to 20 cm depth. The ridges in the sampled ridge and furrow complex at that site are orientated orthogonally to the isohypses (Bergmann, 1991), promoting water erosion through the furrows. Thus, topsoil material at furrows may have been carried away. This hypothesis is supported by the high silt content (above 80%) at the site (Table A5.1) and the below ground claystone causing waterlogging, as confirmed through a high proportion of redoximorphic features in the horizons above the claystone (Fig. 4.2). It is also possible that at the Eddessen site, temporal waterlogging is higher at furrows than at ridges thus improving relative growth conditions for trees. However, this is not likely to have resulted in higher litter or root C input at ridges because the canopy diameter of the beech trees is approximately 10 m, and thus larger than the mean horizontal distance between ridge and furrow of 9 m (Table 4.2). It has also been observed that root density does not decrease with increasing distance to trees (Angst et al., 2016a).

Preservation of buried topsoil SOC under ridge

Despite the SOC content at 0 to 20 cm depth being higher at reference and furrow positions than at ridges, most likely due to higher C input, SOC content was not significantly lower in the buried topsoil horizon at ridges than in corresponding depth at reference and furrow positions (Fig. 4.3 and 4.4). If the C input at the surface is the main driver for SOC storage, not only in topsoils but also in subsoils, the similar SOC content in the buried topsoil layer at ridge, reference and furrow positions implies that a certain amount of SOC in the buried topsoil was preserved over centuries.

In forests of New England, USA, former ploughing horizons at 5 to 20 cm depth have been reported to be still visually identifiable after a century (Foster et al., 2003). In colluvial arable soils of Belgium, 50% of SOC buried through erosion to 25 to 100 cm depth is reported to be still preserved 250 to 300 years after burial (Wang et al., 2014). In two raised fields of the Bolivian Lowlands, a buried fossil Ap horizon at 220 to 230 cm depth has been shown to contain two- to six-fold more SOC than the subsoil horizons directly above it (Rodrigues et al., 2015).

In our study, topsoil burial was much older, because the ridge and furrow system was established at least 691 to 1011 years ago, which is the time when the villages at the study sites were first mentioned (Table 4.2). It is possible that ridge and furrow systems are even older and have remained at fixed positions over centuries (Bartussek, 1982). The buried soil was below the regular plough layer at least since village abandonment (approximately between 115 and 358 years ago; Table 4.2).

In addition to the great age of the buried topsoil, it has to be taken into account that the burial process in ridge and furrow fields might have occurred slowly. Topsoil originating from the furrows was deposited through ploughing in the same direction on top of the ridges. With continued cultivation, subsoil from the furrows was also moved to the ridges. The relatively shallow depth of the buried topsoil layer and the fact that it was not buried in one single operation has to be considered when comparing it with buried topsoil in other studies.

Buried SOC was not completely mineralised and can be considered stable on a long-term basis. On the one hand, the proportion of total SOC in HF was highest in buried topsoil compared with all depths at reference positions (Fig. 4.8). On the other hand, narrower C/N ratios in buried topsoil at ridges than in the same depth at reference and furrow positions (Fig. A5.4) indicate a higher degree of SOC transformation.

4.6 Conclusions

Medieval ridge and furrow cultivation was found to have modified SOC storage even centuries after natural reforestation. The primary SOC effect from the legacy of excavating furrows and building up ridges by ploughing was an uneven spatial distribution of SOC stocks, with higher SOC stocks at furrows. This SOC distribution pattern could be explained by differences in C input especially from leaf litter. We also observed long-term preservation of SOC from former topsoil buried under ridges. Topsoil layers that were buried 700 to 1000 years ago were still visible in ridge and furrow fields and contained more SOC at four of the five ridge and furrow sites studied. However, topsoil burial did not increase total SOC stocks. We conclude that in forests on land with historical use of a ridge and furrow system, recent processes with spatial heterogeneous C input determine SOC distribution, but not necessarily also SOC stocks.

5 Synthesis and general conclusions

In this thesis, burial of soil organic matter was studied by assessing deep-ploughed arable and forest soils as well as medieval ridge and furrow fields under current forest vegetation. Both land management practices entailed a certain soil profile modification, translocating former topsoil material to a greater depth and former subsoil material to the soil surface. Thus, both subsoils and topsoils were drastically altered. The impact of the active burial of large amounts of SOC on long-term C sequestration was found to depend on two main key aspects: (1) the stability against complete mineralisation of buried SOC in subsoil and (2) the additional SOC enrichment in the 'newly formed topsoils'.

5.1 Stability of soil organic carbon in buried topsoils

Through deep ploughing, former topsoils were buried to 55-90 cm depth. In this thesis, the effect of burial on SOC stability was assessed by (i) comparing the SOC content in the buried topsoil stripes with that in the reference topsoil and (ii) through one-year incubation experiments, which enabled comparison of SOC turnover in buried topsoil stripes and reference topsoil under standardised laboratory conditions. The fact that at all sites studied SOC content in the buried topsoil was higher than that in adjacent subsoil stripes and in reference subsoil was a clear sign of a certain level of preservation of buried SOC.

The comparison of SOC content in the buried topsoil stripes with that in the reference topsoil showed very heterogeneous results. At three out of the five sandy cropland sites studied, SOC content in the buried topsoil stripes was similar or even higher than that in the reference topsoil (sites Ahlhorn, Essemühle and Eickenrode, see Chapter 2). For the Ahlhorn and Essemühle sites, data of the SOC content in the topsoil prior to deep ploughing was available in the report on experimental site establishment (Foerster, 1974). Comparing the initial SOC content (non-buried topsoil) with the current SOC content in the buried topsoil stripes, we recorded higher SOC concentrations in the buried topsoil than in the topsoil before deep ploughing, indicating uncertainty in the historical data, but also very small or no changes in the SOC content. On the other hand, at the Elze and Hemmelsberg sites, SOC content in the buried topsoil stripes was 32% and 11%, respectively, lower then that in the reference topsoil. Considering the SOC content in topsoil before deep ploughing, 61% and 73% of the initial SOC were preserved in subsoil since SOC burial after 47 and 36 years, respectively.

In contrast, at all loamy cropland sites studied, SOC content in the buried topsoil stripes was 35-52% lower than that in the reference topsoil. At these sites, between 47 and 60% of the initial SOC before deep ploughing was maintained in buried topsoil stripes after 48 years (see

Chapter 2). These results point out that also at loamy sites SOC contents did not decline until reaching subsoil-typical values. Instead, a “basal SOC content” was reached. Similar trends were observed at three of the four forest sites studied, where SOC content in the buried topsoil stripes was 20-95% lower than that in the reference topsoil (see Chapter 3). Unfortunately, no data was available on SOC content of the forest topsoil before deep ploughing. Thus, it is possible that the SOC loss is overestimated at the forest sites. At the forest site Schwenow, SOC content in the buried topsoil stripes was similar or even higher than that in the reference topsoil similarly to the observations at sandy cropland sites.

The high SOC stability at many of the sandy sites studied was suggested to be related to the land use history as heathland or peatland (Table 4.1). Former heathland soils have been observed to contain very stable SOC possibly related to a high content of hydrophobic and toxic substances for decomposers(Springob and Kirchmann, 2002; Jalal and Read, 1983; Sleutel et al., 2008). On the other hand, it was surprising that SOC preservation in buried topsoil was relatively low at the loamy cropland sites studied because SOC stabilisation has often been found to be related particularly to fine-textured soils (Flessa et al., 2008; Doetterl et al., 2015a). Further research in this regard is needed to confirm if high SOC content and a high SOC stabilisation degree in sandy soils are solely related to the land use history. This is particularly important for regions where these two aspects are found, such as northwestern Germany.

At all sites studied, mineralisable SOC fraction during laboratory incubation was lower in buried topsoil than in reference topsoil (see Chapter 3). Chapter 2 shows differing results for the loamy cropland sites, possibly explained by the shorter incubation time of this part of the study (24 hours compared with a one-year batch incubation used in the latter study presented in Chapter 3). Laboratory incubation under standardised conditions (22°C, 60% water-holding capacity and aeration) turns off possible SOC mineralisation limitations related to low temperature, low oxygen concentrations or soil moisture in the incubated samples. On the basis of these results, it can be concluded that the stability of buried SOC is not solely caused by environmental conditions at greater soil depth. At the moment, laboratory and field observations show inconsistent results in this question related to the great heterogeneity in subsoil resulting in large data variability (C. Rumpel et al., 2012). However, possible SOC mineralisation limitations in subsoil in contrast to topsoil conditions may be due to lack of regular physical disturbance via tillage or intense bioturbation or low availability of nutrients, most importantly N and P (Niebuhr et al., 2016; Fierer et al., 2003).

The influence of site characteristics (climate and management variables) and soil properties on the effect of burial on SOC mineralisation assessed in the present thesis did not reveal the importance of certain key factors for SOC stability. Nevertheless, it was observed that the amount of microbial biomass in buried topsoil was lower than in reference topsoil. High SOC contents and low amounts of microbial biomass have also been observed in buried colluvial topsoils (Struecker and Joergensen, 2015). However, SOC mineralisation rates relative to the amount of microbial biomass were not lowered as revealed by higher metabolic quotients in buried topsoil than in reference topsoil (see Chapter 2). Thus, C_{mic} content *per se* was found not be directly related to SOC mineralisation as has been observed in previous studies (Agnelli et al., 2004; Jörgensen et al., 2002; Lavahun et al., 1996). Wordell-Dietrich et al. (2016) hypothesises this missing relation with the lower carbon use efficiency of microorganisms in subsoil, i.e. organisms incorporate substrate into their biomass to a lower degree, but need to mineralise large parts of the available substrate to meet their energy demands for growth and maintenance.

Because preservation of high SOC content and higher SOC stability in the buried topsoil than in the reference topsoil was not related to depletion of the fLF (see Chapter 3), commonly defined as the most labile SOC fraction (Lützow et al., 2007), selective preservation of an intrinsic more stable fraction, such as the HF, and concomitant selective mineralisation of a more labile fraction could not fully explain the observations of this thesis. In addition to the partly very high fLF contents in buried topsoil, a comparison with the SOC mineralisation in long-term bare fallow experiments (see Chapter 2) revealed that long-term SOC mineralisation rates were lower in buried topsoils than in bare fallow topsoils. Thus, it can be concluded that. These observations point out the importance of environmental conditions at greater soil depth that require more precise investigation.

Radiocarbon analyses of SOC fractions were performed in the frame of this thesis in order to gain insight into the turnover times of SOC in buried topsoil and reference topsoil. This would provide direct evidence if SOC turnover is slowered due to burial. Results showed higher $\Delta^{14}C$ values in buried topsoil than in reference topsoil indicating that in buried topsoil C exchange with the atmosphere has been slow and that significant radioactive decay has occurred, i.e. input of fresh carbon into these soil fractions was drastically reduced due to burial. In order to model SOC turnover times and make a proper comparison of buried topsoil with reference topsoil, ^{14}C measurements of two, or better three, different years would be necessary. An additional challenge to understanding C turnover in buried soils are the uncertainties regarding the magnitude of fresh C input. This input at greater soil depth is also possibly influences by certain lag times, i.e. a year difference between CO_2 fixation in plants and plant litter reaching the subsoil (Schrumpf and K. Kaiser, 2015).

A high degree of stability was also confirmed for SOC in the buried topsoil at former ridge and furrow fields established at least 691 to 1011 years ago (see Chapter 4), highlighting that SOC burial leads to long-term SOC sequestration. Buried SOC in these systems was characterised by lower *in situ* mineralisation rates (Fig. 9 in Chapter 4) and higher HF content (Fig. 8 in Chapter 4) than the reference topsoil. These results would support the theory of selective preservation as the mechanisms for SOC mineralisation. Comparing topsoil burial in ridge and furrow systems with topsoil burial in deep ploughed soils must, however, be conducted under certain considerations. The burial process in ridge and furrow fields probably occurred slowly. The separation of buried topsoil and overlying subsoil was not sharp due to continuous cultivation over decades to centuries, in which subsoil from the furrows was ploughed toward the ridges. This might explain that most of the buried SOC under ridges was mineral associated as it is typical for subsoil (Eusterhues et al., 2003).

5.2 Accumulation of soil organic carbon in newly established topsoils

Deep ploughing is a land management operation, in which soil horizons are rotated by a certain angle, often 130°, thus causing a drastic profile modification. As a consequence, stripes of topsoil and subsoil horizons alternate from the surface down to deep ploughing depth. The proportion of each horizon stripe depends on former horizon mightiness, rotation angle, speed of the deep mouldboard plough, soil moisture, soil coherence mostly dependant on soil texture, among others (e. Deutsches Institut für Normung, 2015). With continuous, subsequent plant growth on top of the deep ploughed soil, SOC is accumulated and a 'new' topsoil is established over time.

At the sites studied, significant differences in topsoil SOC stocks between the non-deep-ploughed reference and deep-ploughed soil persisted even 48 years after the deep ploughing event, although a trend was visible toward lower SOC deficit with increasing number of years since deep ploughing (see Chapters 2). An unexpected result was that forest topsoil showed a slower SOC accumulation than cropland topsoil in spite of litter layer SOC stocks being similar (see Chapter 3). These observations may be explained by the following:

- Through regular tillage operations, crop residues and leaf litter are directly incorporated into the mineral soil. On the contrary, aboveground forest litter first ends up in the forest floor and is thereafter slowly and not completely transferred into the mineral soil (Fisher and Binkley, 2013).
- At the forest sites studied, topsoil was more acidic than in cropland (3-5 compared to 5-7) and coniferous trees, i.e. *Pinus sylvestris*, produces litter that is decomposed more slowly, e.g. because of a higher lignin content than deciduous vegetation (Berg and McClaugherty, 2014).
- The maximum possible SOC sequestration in topsoil of deep-ploughed forests is restricted by the available N supply, whereas N restriction is less probable in topsoil of deep-ploughed cropland due to a 4- to 15-fold greater N input by mineral and organic fertilisers (see Table 2.2).

These results underscore, moreover, the previous observations that SOC losses may be very quick, but SOC accumulation is regularly slow, as it has been observed that C accumulation occurs much more slowly in afforestations than the carbon loss following deforestation (Poeplau et al., 2011). Beare et al. (2014) have suggested that the rate of possible SOC sequestration is limited by (i) the SOC content already present and (ii) the maximum loading rate of the soil matrix. The latter was related to the specific surface area of soil minerals in relation with pH. With these parameters, a saturation deficit was calculated in order to provide an estimate on which soils were most able to sequester further SOC. One main finding was that the modelled saturation deficit increased with soil depth.

Rasse et al. (2005) have also pointed out that C-free mineral surfaces are the components responsible for accumulation and stabilisation of OC in subsoil. For sandy soils, SOC accumulation might occur mostly as particulate organic matter (Liang et al., 2003) or depend on the poorly crystalline oxides content (Kleber et al., 2005). With all these considerations, the admixture of subsoil material to the topsoil seems promising means of enhancing the maximum SOC sequestration in the topsoil. For example, 'flipped' soils in New Zealand, which were inverted with an excavator down to 1 to 3 m depth, showed a five-fold increase of SOC content within eight years since subsoil material was brought up to the surface (Thomas et al., 2007). Also, continuous SOC accumulation in topsoils was observed after plough horizon deepening from 25 to 35 cm, through which subsoil material was mixed with former topsoil, in many European countries during the 1960s and 1970s (Nieder and Richter, 2000). Nevertheless, it must be taken into account that initial lowering of SOC content in the topsoil might entail certain fertility losses for several years or even decades.

At ridge and furrow sites studied, SOC accumulation in forest topsoil was shaped by the microtopography because litter preferentially accumulated at furrow positions causing an enhanced C input compared with the ridge field positions. This resulted in higher SOC stocks at furrows than at ridges. Because furrow positions were intensely excavated, it can be assumed that after abandonment of ridge and furrow cultivation (i.e. agricultural land use), exactly these positions had the highest proportion of subsoil material in the upper cm of the soil. Therefore, also here, an intense SOC accumulation took place over centuries driven by the relatively high C input.

5.3 Further relevant aspects for climate change mitigation

An evaluation of the effect of carbon burial on SOC sequestration and thus its potential as a measure for climate chance mitigation can only be complete when taking its effect on crop and forest yields into consideration. A meta-analysis on the effect of deep tillage on crop yields indicates that especially at sites with root restrictive layers, such as hardpans in Bhs-horizons of Podzols, yield response to deep tillage is mostly positive (Schneider et al., 2017). However, long-term observations are very rare and thus it is highly uncertain how enduring the effect of deep ploughing on improving plant growth conditions may be. The decreased topsoil fertility in the first years deep tillage might additionally cause a higher fertilizer demand, which would lower the climate change mitigation effect of deep ploughing.

Moreover, it could be objected that the required fossil fuel amount to conduct a carbon burial operation such as deep ploughing is too high. Assuming a diesel demand of 300 L ha^{-1} (as was informed by an agricultural machinery company) for deep ploughing and a CO_2 emission factor of 2.67 emitted kg CO_2 L^{-1} diesel (E. Deutsches Institut für Normung, 2013), it can be calculated that the deep ploughing operation leads to the emission of 0.8 Mg CO_2 ha^{-1}. The average SOC sequestration observed in cropland soils was 42 Mg ha^{-1}. Thus, the fuel demand represents only 2% of the average stored SOC in cropland soils. Furthermore, nitrous oxide and methane may possibly be emitted during the deep ploughing operation, but no quantification in this regard is available.

5.4 Perspectives on future research

Topsoil burial has proven to be an effective measure of SOC sequestration in subsoil. To further elucidate the mechanisms of SOC stabilisation in buried topsoils and eventually develop a conceptual model that will improve the understanding of SOC stabilisation in top- and subsoil, the following hypothesis should be tested:

- Exposure of buried topsoil samples to physical disturbance in the form of tillage, drying and rewetting and freezing-thawing cycles promotes SOC redistribution, thus enhancing the probability of microorganisms and SOC to meet and results in higher SOC mineralisation.
- Concentrations of available forms of nutrients, such as N and P are lower in buried topsoil than in reference topsoil and similar to that in reference subsoil. Nutrient amendments to buried topsoil result in higher SOC mineralisation compared to non-amended treatments.
- Soils with a historical land use as heathland or that were historically peatland soils still contain large amounts of hydrophobic organic matter, which shows high contact angles.
- Repeated measurements of ^{14}C and long-term litter input data collection along with a more precise estimation of C input via root litter can help elucidate turnover times of buried SOC compared with topsoil SOC.

- A greater insight into the role of SOC inputs on stabilisation or destabilisation of buried SOC could be gained when combining topsoil burial experiments with bare fallow treatments.

Regarding the aspect of SOC sequestration in 'newly established topsoil' of deep ploughed soil, the role of a possible N limitation could be assessed comparing treatments with and without mineral and\or organic N fertilisation in deep ploughed forest soils. In the context of climate change adaptation measures, buried topsoil stripes may enhance the resilience of plant growing conditions due to higher organic matter content providing nutrients and water retention, thus mitigating the predicted increased drought stress in climate change scenarios.

A Appendix A - Deep ploughing increases agricultural soil organic matter stocks

Table A3.1: SOC and N stocks in the deep ploughed and reference subplots. SD (standard deviation), SE (standard error).

				SOC stock (Mg ha^{-1})			N stock (Mg ha^{-1})		
Site	**Subplot**	**Substrate**	**n**	**Mean**	**SD**	**SE**	**Mean**	**SD**	**SE**
AH	deep ploughed	sandy	5	450.72	104.38	46.68	16.96	3.23	1.44
AH	reference	sandy	5	161.85	28.74	12.85	9.74	0.61	0.27
BT	deep ploughed	loamy	5	74.95	7.46	3.34	9.65	0.59	0.26
BT	reference	loamy	5	63.17	6.51	2.91	9.24	1.36	0.61
DB	deep ploughed	loamy	5	85.23	5.74	2.57	9.27	2.58	1.15
DB	reference	loamy	5	88.43	7.69	3.44	5.56	0.34	0.15
EM	deep ploughed	sandy	5	148.69	21.45	9.60	11.07	1.78	0.79
EM	reference	sandy	5	121.79	17.32	7.74	8.43	0.98	0.44
ER	deep ploughed	sandy	5	151.35	28.87	12.91	11.28	1.98	0.89
ER	reference	sandy	5	113.66	11.97	5.35	8.53	0.43	0.19
EZ	deep ploughed	sandy	5	78.40	6.78	3.03	8.27	1.38	0.62
EZ	reference	sandy	5	67.84	14.6	6.53	8.46	0.77	0.34
HB	deep ploughed	sandy	5	222.52	33.09	14.80	10.23	0.72	0.32
HB	reference	sandy	5	215.50	16.78	7.51	6.61	0.88	0.39
HT	deep ploughed	loamy	5	79.08	12.26	5.48	6.28	0.53	0.24
HT	reference	loamy	5	61.77	3.56	1.59	4.89	1.13	0.51
SZ	deep ploughed	loamy	5	76.52	5.95	2.66	12.07	1.89	0.85
SZ	reference	loamy	5	57.92	8.21	3.67	6.90	0.42	0.19
WB	deep ploughed	loamy	5	72.32	7.37	3.29	9.62	2.78	1.24
WB	reference	loamy	5	62.76	6.74	3.01	8.22	0.66	0.29

Table A3.2: SOC concentrations before deep ploughing.

Values of depth increment 0-25 cm for site AH were too low to be realistic. Thus, we instead used the values measured in 2014 in the reference topsoil. This was also the case for the SOC concentration below 70 cm, since none was provided in (Foerster, 1974).

Site	Reference	Sampling depth (cm)	SOC before deep ploughing (g kg^{-1})
AH	(Foerster, 1974)	0-25 (2014)	3.61
		30-50	0.76
		50-70	0.4
		58-90 (2014)	0.21
DB	(Scheffer & Meyer, 1970)	0-20	2.02
		20-45	1.36
		45-90	0.67
EM	(Foerster, 1974)	0-20	2.21
		30-50	0.95
		50-70	0.25
EZ	(Reichenbach, 1972)	0-25	1.45
		25-45	0.58
		45-75	0.35
HT	(Czeratzki, 1968)	0-30	1.1
		30-50	0.4
		50-80	0.2
SZ	(Czeratzki, 1968)	0-30	1.2
		30-50	0.5
		50-80	0.5

Table A3.3: SOC stocks in deep ploughed topsoils

SOC contents directly after deep ploughing were retrieved from the studies below the table or an approximation was calculated. Bulk density changes through deep ploughing were assumed to be levelled off after an initial settling phase of the deep ploughed soil. Thus, we assumed no differences in bulk density. Values measured in 2013/2014 are provided as arithmetic means ± standard errors (n=5). The calculation of SOC accumulation rate did not consider changes of the topsoil depth.

	SOC (g kg^{-1})		Bulk density (g cm^{-3})	Topsoil depth (cm)	SOC stocks (Mg ha^{-1})		Years between deep ploughing and sampling	SOC accumulation (Mg ha^{-1} yr^{-1})
Site	directly after deep ploughing	measured in 2013/2014	measured in 2013/2014		directly after deep ploughing	measured in 2013/2014		
AH	13.4[e]	29.4±6	1.28±0.06	25	42.9	94.1	47	1.09
BT	4.6[a]	9.1±0.3	1.41±0.05	32	20.8	41.1	48	0.42
DB	7.6[e]	9.3±0.4	1.45±0.05	32	35.3	43.2	48	0.16
EM	11.1[e]	12.7±0.8	1.47±0.06	35	57.1	65.3	46	0.18
EZ	9.2[e]	10.4±0.4	1.37±0.03	27	34.0	38.5	47	0.09
HB	20.3[c]	25.6±2.6	1.35±0.02	38	104.1	131.3	36	0.75
HT	6.3[e]	9.9±0.4	1.50±0.03	30	28.4	44.6	48	0.34
SZ	7.1[e]	9.6±0.2	1.28±0.05	32	29.1	40.6	47	0.25
WB	4.6[d]	7.4±0.2	1.50±0.05	35	24.2	38.9	48	0.31
								Mean = 0.40
								SE = 0.11

[a] (Grosse, 1974), [b] (Scheffer & Meyer, 1970), [c] (Bartels & Pluquet, 1989), [d] (Grosse & Renger, 1974),
[e] own approximation as

$$SOC\ deep\ ploughed\ topsoil = \sum_{i=1}^{n} \left(SOC\ before\ deep\ ploughing\ \times \frac{\rho_i \times D_i}{\sum_{i=1}^{n} (\rho_i \times D_i)} \right)$$

where ρ is the bulk density, SOC before deep ploughing from Table S2 and D the length of the depth increment over n depth increments down to deep ploughing depth.

Table A3.4: Soil properties from the topsoils of the reference subplots assessed to identify factors influencing the effect of deep ploughing on SOC stocks. Sites AH and SZ had no depth increment 4 because deep ploughing horizon ranged to 90 cm depth.

site	Potential SOC mineralization rate ($\mu g\ CO_2$-C g^{-1} soil h^{-1})	specific potential SOC mineralization rate ($\mu g\ CO_2$-C g^{-1} SOC h^{-1})	Microbial Biomass ($\mu g\ C_{mic}\ g^{-1}$ soil)	specific C_{mic} (mg $C_{mic}\ g^{-1}$ SOC)	qCO_2 ($\mu g\ CO_2$-C $mg^{-1}\ C_{mic}\ h^{-1}$)	pH: depth increment 1	pH: depth increment 2	pH: depth increment 3	pH: depth increment 4	CEC ($mmol_c\ kg^{-1}$ soil)	Fe_{D-O} (g kg^{-1} soil)	$Fe+Al_O$ (g kg_{-1} soil)	Sand (%)	Silt (%)	Clay (%)	N input from mineral fertilizers (kg ha^{-1} yr^{-1})	N input from organic fertilizers (kg ha^{-1} yr^{-1})	C input from organic fertilizers (kg ha^{-1} yr^{-1})
AH	0.3	9	221	6	1.5	5.6	5.3	5.2	-	91.3	0.0	5.6	86	10	4	79	170	986
BT	0.3	32	368	37	0.9	6.5	6.5	6.6	6.7	96.8	3.4	3.0	5	82	13	201	0	0
DB	0.3	23	210	18	1.3	6.8	6.8	6.6	6.7	120.8	4.7	2.8	3	82	15	163	94	475
EM	0.2	23	114	11	2.0	4.6	4.5	5.3	5.4	22.4	1.2	3.2	88	8	4	128	153	550
ER	0.4	17	204	8	2.1	5.8	5.7	5.7	5.7	113.8	1.2	3.9	87	8	5	126	116	1 159
EZ	0.2	14	133	11	1.3	5.4	5.4	5.5	4.9	41.0	0.9	2.0	84	12	4	121	27	60
HB	0.5	15	306	9	1.7	5.4	4.8	4.7	4.7	78.7	0.4	1.2	94	3	3	77	171	375
HT	0.2	22	255	26	0.9	6.3	6.6	6.6	6.5	106.1	4.0	2.7	3	83	14	185	91	405
SZ	0.4	37	285	24	1.6	6.8	6.9	7.1	-	123.4	3.8	3.1	3	83	14	123	34	46
WB	0.2	25	236	30	0.9	5.6	6.6	6.5	7.0	107.0	4.2	4.5	3	80	17	182	11	15

Table A3.5: Model outputs for assessment of effect of pedogenic properties on SOC stock increase through deep ploughing (only significant effects shown). SD (standard deviation), SE (standard error), df (degrees of freedom).

Random effects		
Formula	~1|substrate	~1 | site within substrate
SD of Intercept	46.1	38.6
SD of Residual	-	6.7

Variance function with the structure: different SD per stratum (Formula: ~1 | tillage * substrate)				
Parameter	reference*loamy	deep ploughed*loamy	reference*sandy	deep ploughed*sandy
Estimate	1.0	1.3	2.7	19.7

Fixed effect: cstock ~ tillage (deep ploughed vs. reference)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	101.2	34.8	89	2.9	<0.01
Tillage deep ploughed	10.3	2.2	89	4.7	<0.01

Fixed effect: cstock ~ tillage * fed.min.o (Fe_{D-O})					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	60.7	63.7	87	0.9	0.3
Tillage deep ploughed	53.1	16.6	87	3.2	<0.01
fed.min.o	17.1	4.0	87	4.2	<0.01
Tillage deep ploughed : fed.min.o	-12.8	3.8	87	-3.4	<0.01

Fixed effect: cstock ~ tillage * silt (Silt content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	143.9	18.3	87	7.9	<0.01
Tillage deep ploughed	74.9	25.1	87	3.0	<0.01
silt	-95.3	30.6	87	-3.1	<0.01
Tillage deep ploughed : silt	-86.3	31.8	87	-2.7	<0.01

Fixed effect: cstock ~ tillage * sand (Sand content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	-129.5	228.7	87	-0.6	0.6
Tillage deep ploughed	6.4	2.7	87	2.4	0.02
sand	506.2	230.0	87	2.2	0.03
Tillage deep ploughed : sand	70.1	25.6	87	2.7	<0.01

Fixed effect: cstock ~ tillage * clay (Clay content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	67.1	56.9	87	1.2	0.2
Tillage deep ploughed	45.1	19.4	87	2.3	0.02
clay	366.2	169.2	87	2.2	0.03
Tillage deep ploughed : clay	-263.9	129.7	87	-2.0	0.04

Table A3.6: Soil properties from the buried topsoils of the deep ploughed subplots assessed to identify factors influencing potential SOC mineralization rates

Site	Specific potential SOC mineralization rate (μg CO_2-C g^{-1} SOC h^{-1})	Microbial Biomass (μg C_{mic} g^{-1} soil)	Specific C_{mic} (mg C_{mic} g^{-1} SOC)	qCO_2 (μg CO_2-C mg^{-1} C_{mic} h^{-1})	pH	CEC ($mmol_c$ kg^{-1} soil)	Fe_{D-O} (g kg^{-1} soil)	$Fe+Al_O$ (g kg^{-1} soil)	Sand (%)	Silt (%)	Clay (%)
AH	0.02 (±0.001)	36.3 (±1.15)	5.8 (±0.2)	3.7 (±0.2)	3.4	8.4	0.2	1.2	91	3	6
BT	0.26 (±0.028)	55.5 (±0.96)	120.8 (±2.1)	2.2 (±0.2)	6.9	9.2	4.4	3.3	5	82	13
DB	0.24 (±0.010)	52.0 (±0.30)	94.5 (±0.6)	2.5 (±0.1)	6.3	1.9	3.6	2.7	4	84	12
EM	0.07 (±0.005)	55.0 (±0.31)	24.6 (±0.1)	2.8 (±0.2)	5.4	4.8	1.3	5.1	75	20	4
ER	0.04 (±0.009)	12.7 (±5.35)	4.9 (±2.1)	13.6 (±6.6)	5.0	5.2	0.2	1.8	90	7	3
EZ	0.06 (±0.007)	32.0 (±0.88)	40.0 (±1.1)	1.6 (±0.2)	5.2	3.3	1.4	1.7	85	12	4
HB	0.04 (±0.003)	36.5 (±0.77)	12.9 (±0.3)	3.1 (±0.2)	4.1	4.5	0.3	1.3	95	3	3
HT	0.66 (±0.016)	64.9 (±3.58)	101.4 (±5.6)	6.5 (±0.2)	7.2	10.7	3.7	3.1	4	83	13
SZ	0.25 (±0.010)	75.9 (±1.08)	118.6 (±1.7)	2.2 (±0.1)	7.3	11.5	4.8	2.7	2	84	14
WB	0.48 (±0.017)	111.2 (±7.36)	218.1 (±14.4)	2.2 (±0.1)	6.8	9.7	3.4	3.2	2	85	13

Table A3.7: Model outputs for assessment of the specific potential SOC mineralization in buried topsoils of the deep ploughed subplots (only significant effects shown). SD (standard deviation), SE (standard error), df (degrees of freedom).

Random effects	
Formula	~1|site
SD of Intercept	0.2
SD of Residual	0.03

Fixed effect: specificSOCmineralization ~ 1					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	0.2	0.1	26	3.1	<0.01

Fixed effect: specificSOCmineralization ~ pH (pH)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	-0.5	0.21	26	-2.5	0.02
pH	0.1	0.04	8	3.6	<0.01

Fixed effect: specificSOCmineralization ~ silt (Silt content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	0.01	0.07	26	0.1	0.9
silt	0.4	0.11	8	4.0	<0.01

Fixed effect: specificSOCmineralization ~ sand (Sand content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	0.4	0.06	26	6.4	<0.01
sand	-0.4	0.10	8	-4.0	<0.01

Fixed effect: specificSOCmineralization ~ clay (Clay content)					
Parameter	Value	SD	df	t-value	p-value
(Intercept)	-0.1	0.09	26	-1.0	0.3
clay	3.5	0.89	8	4.0	<0.01

B Appendix B - Stability of buried carbon in deep ploughed forest and cropland soils - implications for carbon stocks

Table A4.1: Topsoil C and N stocks difference in reference and deep ploughed soils (N=5 determined from soil coring). Absolute stock differences represent the subtraction of stocks in deep ploughed topsoils minus those in reference topsoils. Thus, positive values represent higher stocks in topsoil of the deep ploughed soils. Relative differences were calculated by dividing the stock in the topsoil of the deep ploughed by the stock in the reference topsoil. Values higher than 1 represent that the deep ploughed soils had higher stocks.

Land use	Site	Absolute C stock difference [Mg ha^{-1}]	Relative C stock difference [Mg ha^{-1}]	Absolute C stock difference relative to years since deep ploughing [Mg ha^{-1} yr^{-1}]	Absolute N stock difference [Mg ha^{-1}]	Relative N stock difference [Mg ha^{-1}]	Absolute N stock difference relative to years since deep ploughing [Mg ha^{-1} yr^{-1}]
Forest	Lindenburg	-23.8±6.7	0.4	-0.6	-0.7±0.2	0.5	-0.02
Forest	Rebberlah	-3.4±2.7	0.9	-0.1	-0.3±0.1	0.7	-0.01
Forest	Schwenow	-8.5±3.0	0.7	-0.2	-0.1±0.1	0.9	-0.00
Forest	Viborg	-8.8±2.6	0.6	-0.4	-0.5±0.1	0.6	-0.02
Cropland	Elze	-8.5±2.9	0.8	-0.2	-0.6±0.2	0.8	-0.01
Cropland	Essemühle	-11.6±6.6	0.9	-0.3	1.0±0.7	1.2	0.00
Cropland	Hemmelsberg	-23.6±15.9	0.9	-0.7	-1.8±0.8	0.8	-0.05
Cropland	Banteln	0.1±2.7	1.0	0.0	0.1±0.2	1.0	0.00
Cropland	Drüber	-6.1±3.0	0.9	-0.1	-0.6±0.3	0.9	-0.01
Cropland	Halchter	0.3±1.2	1.0	0.0	0.1±0.2	1.0	0.00
Cropland	Salzgitter	0.5±2.5	1.0	0.0	-0.2±0.7	1.0	0.00
Cropland	Warberg	-2.7±2.7	0.9	-0.1	0.03±0.3	1.0	0.00

Table A4.2: Characteristics of deep ploughed forest and cropland study sites. MAT: mean annual temperature, MAP: mean annual precipitation

Site	Latitude, Longitude	MAT [°C]	MAP [mm]	Elevation [m above mean sea level]
Lindenburg	52°32'39“ N 9°34‘15“ E	9.2	700	64
Rebberlah	52°43‘31“ N 10°10.27“ E	8.8	730	88
Schwenow	52°9‘10“ N 14°2‘2“E	9.4	849	63
Viborg	56°24‘1“ N 9°19‘5“E	8.4	801	49
Elze	52°35'06"N 9°45'29"E	9.2	698	38
Essemühle	52°45'51"N 8°28'35"E	9.2	717	30
Hemmelsberg	53°04'60"N 8°19'40"E	9.2	750	10
Banteln	52°05'14"N 9°44'56"E	9.2	703	86
Drüber	51°45'28"N 9°54'22"E	8.9	687	154
Halchter	52°08'45"N 10°30'32"E	9.0	642	54
Salzgitter	52°04'12“N 10°27'18"E	9.1	647	116
Warberg	52°11'09"N 10°54'17"E	8.9	649	141

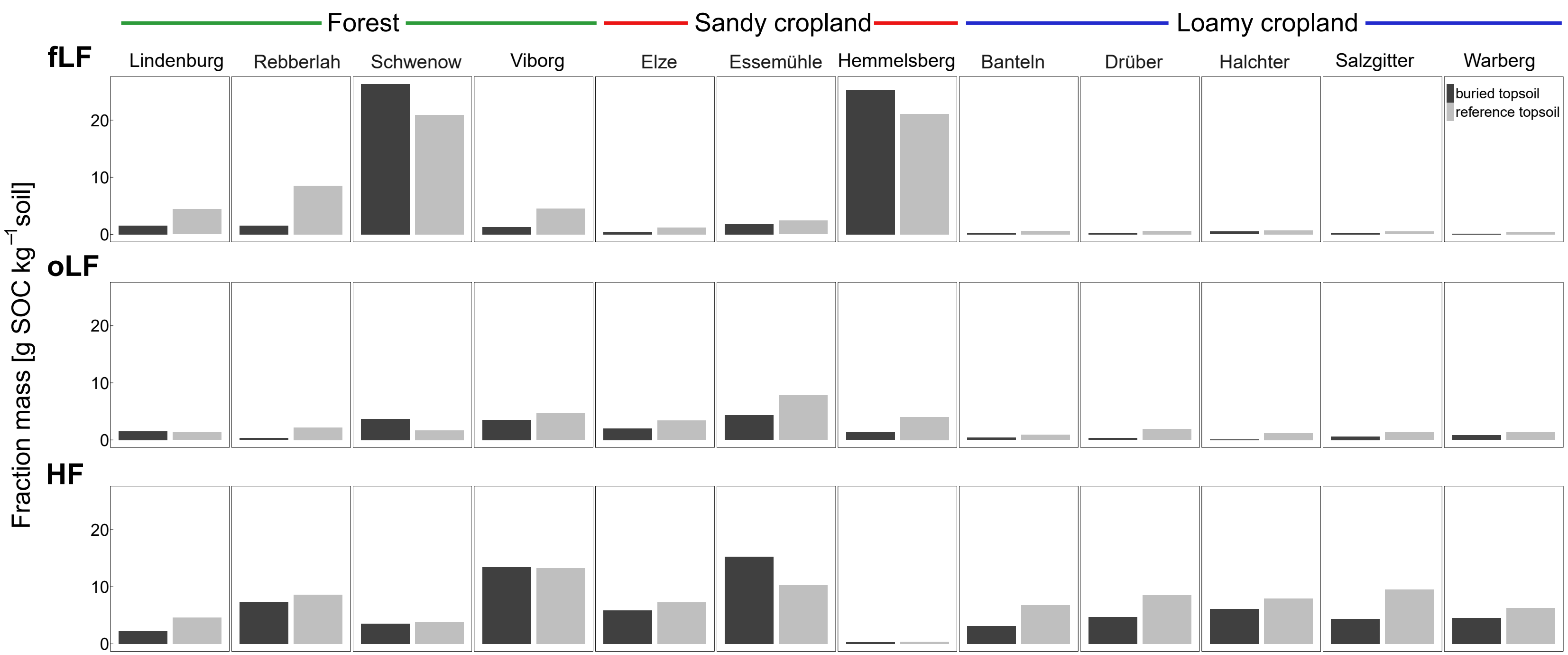

Figure A4.1: Relative mass proportion of bulk soil in SOC fractions (fLF: free light fraction, oLF: occluded light fraction, HF: heavy fraction)

C Appendix C - Legacy of medieval ridge and furrow cultivation on soil organic carbon distribution and stocks in forests

Table A5.1: Soil properties at the ridges and reference position of the five ridge and furrow sites.

Site	Cross section position	Depth increment	SOC [g kg^{-1}]	N_{tot} [g kg^{-1}]	CEC [mmol$_c$ kg^{-1}]	pH	Fe_O+Al_O [g kg^{-1}]	Fe_{D-O} [g kg^{-1}]	Sand [%]	Silt [%]	Clay [%]
Bad Helmstedt	reference	0 - 10 cm	32.1	2.0	9.3	4.6	2.8	3.3	64	27	9
Bad Helmstedt	reference	20 - 44 cm	3.2	0.3	2.2	4.2	2.0	2.8	69	26	5
Bad Helmstedt	ridge	0 -10 cm	26.4	1.6	5.8	4.6	2.8	2.7	67	26	7
Bad Helmstedt	ridge	20 - 44 cm (buried topsoil)	3.6	0.3	1.6	4.2	2.0	2.3	70	23	7
Bad Helmstedt	ridge	44 - 68 cm (non-tilled subsoil)	1.6	0.1	1.6	4.1	1.8	2.2	71	25	4
Eddessen	reference	0 -10 cm	39.4	2.1	5.8	3.1	5.4	2.5	4	83	13
Eddessen	reference	20 - 33 cm	5.5	0.5	3.3	3.8	5.0	2.3	3	85	12
Eddessen	ridge	0 -10 cm	38.4	2.2	4.9	3.3	5.1	2.1	4	82	14
Eddessen	ridge	20 - 33 cm (buried topsoil)	6.0	0.5	2.9	3.8	5.5	2.2	5	82	13
Eddessen	ridge	33 - 46 cm (non-tilled subsoil)	3.5	0.4	3.0	3.9	4.7	2.7	5	83	12
Hohnstedt	reference	0 -10 cm	34.2	1.7	4.0	3.2	5.6	2.4	76	17	7
Hohnstedt	reference	20 - 33 cm	4.7	0.3	1.8	3.7	1.9	0.7	82	14	4
Hohnstedt	ridge	0 -10 cm	18.5	1.0	2.9	3.3	2.1	1.3	76	18	6
Hohnstedt	ridge	20 - 33 cm (buried topsoil)	5.6	0.4	2.3	3.6	2.8	1.1	73	20	7
Hohnstedt	ridge	33 - 46 cm (non-tilled subsoil)	3.5	0.3	1.6	3.8	2.6	0.8	78	17	5
Kreuzheide	reference	0 -10 cm	53.5	2.6	13.6	4.0	1.8	1.3	77	15	8
Kreuzheide	reference	20 - 56 cm	6.3	0.3	2.4	4.1	2.9	1.1	83	13	6
Kreuzheide	ridge	0 -10 cm	36.5	1.7	5.6	4.0	1.5	1.2	81	14	5
Kreuzheide	ridge	20 - 52 cm (buried topsoil)	4.8	0.3	1.8	4.1	2.1	0.8	83	14	4
Kreuzheide	ridge	52 - 84 cm (non-tilled subsoil)	4.2	0.2	1.4	4.3	2.8	1.2	81	15	4
Reinhausen	reference	0 -10 cm	24.2	1.8	8.1	4.8	5.6	4.1	8	76	16
Reinhausen	reference	20 - 43 cm	4.0	0.5	6.7	4.3	5.5	4.3	6	75	19
Reinhausen	ridge	0 -10 cm	18.5	1.3	6.9	4.4	4.6	3	10	75	15
Reinhausen	ridge	20 - 43 cm (buried topsoil)	4.6	0.5	5.2	4.3	5.8	3.3	8	77	15
Reinhausen	ridge	43 - 66 cm (non-tilled subsoil)	2.4	0.4	8.8	4.8	5.3	5.4	6	72	22

Table A.2: Mass corrected soil organic carbon (SOC) stocks down to the maximum depth of the lightest core per ridge and furrow site

Site	Cross section position	SOC stock [Mg ha^{-1}]	Depth of lightest core [cm]	Soil mass [g]
Bad Helmstedt	ridge	53.1	45	5318
Bad Helmstedt	middle 1	47.8	45	5318
Bad Helmstedt	reference	47.1	45	5318
Bad Helmstedt	middle 2	53.4	45	5318
Bad Helmstedt	furrow	58.0	45	5318
Bad Helmstedt	middle 2	46.8	45	5318
Bad Helmstedt	reference	39.2	45	5318
Bad Helmstedt	middle 1	51.4	45	5318
Bad Helmstedt	ridge	40.7	45	5318
Bad Helmstedt	middle 1	43.2	45	5318
Bad Helmstedt	reference	47.1	45	5318
Bad Helmstedt	middle 2	51.4	45	5318
Bad Helmstedt	furrow	56.5	45	5318
Bad Helmstedt	middle 2	58.8	45	5318
Bad Helmstedt	reference	49.8	45	5318
Bad Helmstedt	middle 1	51.7	45	5318
Eddessen	ridge	77.7	70	7934
Eddessen	middle 1	83.3	70	7934
Eddessen	reference	69.9	70	7934
Eddessen	middle 2	74.0	70	7934
Eddessen	furrow	67.2	70	7934
Eddessen	middle 2	74.6	70	7934
Eddessen	reference	63.6	70	7934
Eddessen	middle 1	67.3	70	7934
Eddessen	ridge	72.9	70	7934
Eddessen	middle 1	62.0	70	7934
Eddessen	reference	63.5	70	7934
Eddessen	middle 2	59.7	70	7934
Eddessen	furrow	70.5	70	7934
Eddessen	middle 2	68.0	70	7934
Eddessen	reference	70.8	70	7934
Eddessen	middle 1	80.1	70	7934
Hohnstedt	ridge	50.6	62	6429
Hohnstedt	middle 1	55.9	62	6429
Hohnstedt	reference	48.8	62	6429
Hohnstedt	middle 2	66.1	62	6429
Hohnstedt	furrow	73.6	62	6429
Hohnstedt	middle 2	64.0	62	6429
Hohnstedt	reference	53.0	62	6429
Hohnstedt	middle 1	47.9	62	6429
Hohnstedt	ridge	52.3	62	6429
Hohnstedt	middle 1	48.2	62	6429
Hohnstedt	reference	63.0	62	6429
Hohnstedt	middle 2	61.3	62	6429
Hohnstedt	furrow	61.9	62	6429
Hohnstedt	middle 2	47.0	62	6429
Hohnstedt	reference	52.9	62	6429
Hohnstedt	middle 1	50.0	62	6429

Table A.2: Mass corrected soil organic carbon (SOC) stocks down to the maximum depth of the lightest core per ridge and furrow site

Site	Cross section position	SOC stock [Mg ha^{-1}]	Depth of lightest core [cm]	Soil mass [g]
Kreuzheide	ridge	57.6	52	3955
Kreuzheide	middle 1	69.2	52	3955
Kreuzheide	reference	79.3	52	3955
Kreuzheide	middle 2	75.5	52	3955
Kreuzheide	furrow	72.2	52	3955
Kreuzheide	middle 2	82.5	52	3955
Kreuzheide	reference	86.2	52	3955
Kreuzheide	middle 1	55.6	52	3955
Kreuzheide	ridge	55.8	52	3955
Kreuzheide	middle 1	77.1	52	3955
Kreuzheide	reference	62.4	52	3955
Kreuzheide	middle 2	89.7	52	3955
Kreuzheide	furrow	92.9	52	3955
Kreuzheide	middle 2	76.0	52	3955
Kreuzheide	reference	79.5	52	3955
Kreuzheide	middle 1	79.0	52	3955
Reinhausen	ridge	47.2	43	5184
Reinhausen	middle 1	46.8	43	5184
Reinhausen	reference	45.9	43	5184
Reinhausen	middle 2	48.2	43	5184
Reinhausen	furrow	59.1	43	5184
Reinhausen	middle 2	47.6	43	5184
Reinhausen	reference	54.0	43	5184
Reinhausen	middle 1	47.5	43	5184
Reinhausen	ridge	52.4	43	5184
Reinhausen	middle 1	49.1	43	5184
Reinhausen	reference	53.9	43	5184
Reinhausen	middle 2	46.9	43	5184
Reinhausen	furrow	48.5	43	5184
Reinhausen	middle 2	37.4	43	5184
Reinhausen	reference	52.0	43	5184
Reinhausen	middle 1	53.1	43	5184

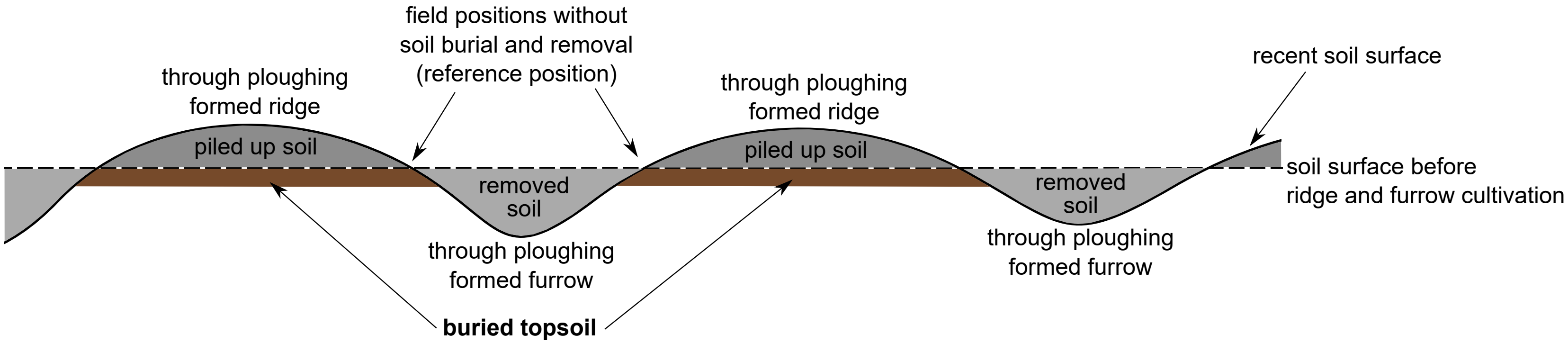

Fig. A5.1: Schematic formation of an undulated soil surface through ridge and furrow cultivation.

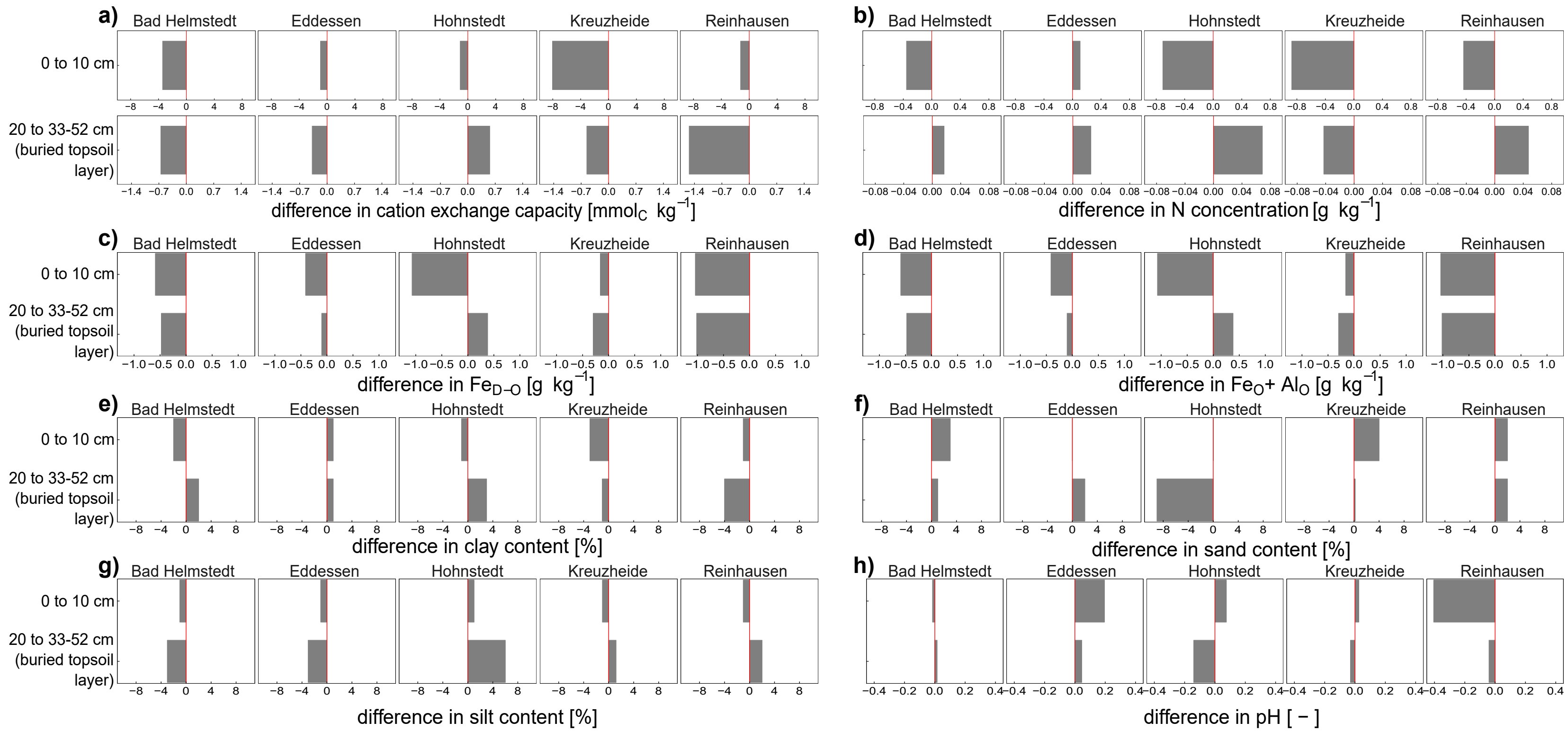

Fig. A5.2: Differences in **a)** cation exchange capacity, **b)** N content, **c)** crystalline iron oxides (Fe_{D-O}), **d)** amorphous iron and aluminium oxides $\overline{(Fe_O+Al_O)}$, **e)** clay content, **f)** sand content **g)** silt content and **h)** pH, ridge minus reference position in 0 to 10 cm depth and in the buried topsoil layer.

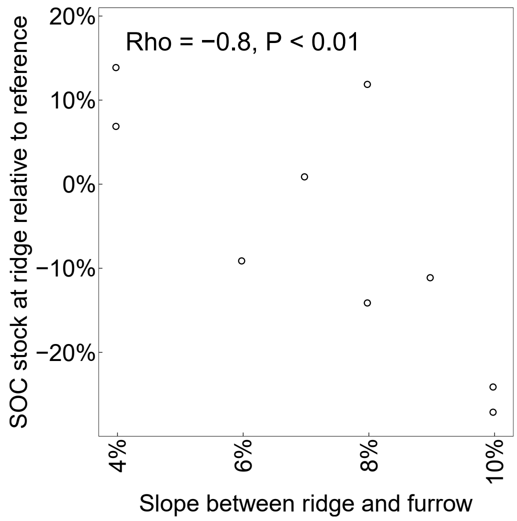

Fig. A5.3: Correlation between SOC stocks at ridges relative to those at reference and the slope between ridge and furrow.

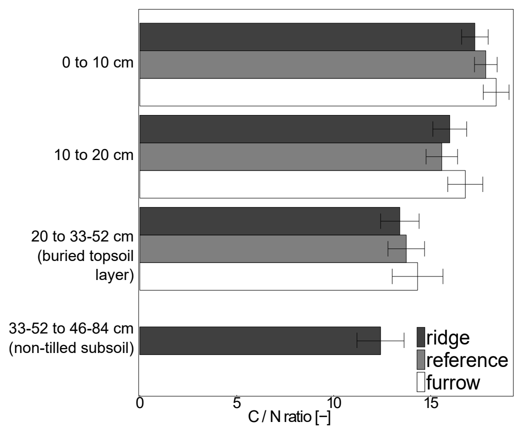

Fig. A5.4: C / N ratios in different depth increments at ridge, reference position and furrow. Arithmetic means from transect core and disturbed profile samples at each site (ridge and furrow: N=15, reference: N=25).

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Alcántara, V., Don, A., Well, R., Nieder, R. (2015). Deep ploughing can increase soil organic carbon stocks through long-term stabilized buried topsoil. 5th International Symposium on Soil Organic Matter, Göttingen, Germany (Poster presentation).

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